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Climate Dynamics

The impact of cold North Atlantic sea surface temperatures on climate: implications for the Younger Dryas cooling (11-10 k)

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Abstract. The sensitivity of global climate to colder North Atlantic sea surface temperatures is investigated with the use of the GISS general circulation model. North Atlantic ocean temperatures 18,000 B.P., resembling those prevalent during the Younger Dryas, were incorporated into the model of the present climate and also into an experiment using orbital parameters and land ice characteristic of 11,000 B.P. The results show that with both 11,000 B.P. and present conditions the colder ocean temperatures produce cooling over western and central Europe, in good agreement with Younger Dryas paleoclimatic evidence. Cooling also occurs over extreme eastern North America, although the precise magnitude and location depends upon the specification of ocean temperature change in the western Atlantic. Despite the presence of increased land ice and colder ocean temperatures, the Younger Dryas summer air temperatures at Northern Hemisphere midlatitudes in the model are warmer than those of today due to changes in the orbital parameters, chiefly precession, and atmospheric subsidence at the perimeter of the ice sheets.

Introduction

The cooling associated with the late-glacial Younger Dryas stade, well defined as a pollen zone in Europe (Jensen 1938; Iversen 1954) and spanning approximately a millenium, followed the warming trend of the Bölling and Alleröd interstades. Geomorphological, lithological, faunal, and floral remains provide substantial evidence for this climatic fluctuation at numerous sites throughout northwestern Europe. Particularly striking are pollen records which indicate that trees which had started growing in response to the climatic warming of the deglaciation were suddenly replaced about 11,000–10,000 B.P. (11–10 k) by shrubs and herbs characteristic of a glacial regime (e.g., Fig. 1). This event is also recorded in lithological and glacial evidence with chronological control. Correlative fluctuations are found in the $^{18}O/^{16}O$ and CO_2 records from the Camp Century and Dye 3 Greenland ice cores (Dansgaard et al. 1982; Oeschger et al. 1984) (Fig. 2a) and in the $^{18}O/^{16}O$ patterns recorded in $CaCO_3$ from lake sediments in France (Eicher et al. 1981) and Switzerland (Oeschger et al. 1980) (Fig. 2b). Large-scale movement of the North Atlantic polar front at this time is indicated by faunal changes in marine cores (Ruddiman and McIntyre 1981). The existence of this fluctuation throughout various paleoclimatic records indicates that the climate system is capable of rapid and extreme changes, at least regionally.

There is no evidence that such extreme climatic oscillations have occured during the past 10,000 years, although the Little Ice Age which western Europe experienced in the eighteenth century might have been a weaker version. However, sharp fluctuations in the $^{18}O/^{16}O$ and CO_2 profiles of the two Greenland ice cores characterize the interval between approximately 25,000 and 40,000 B.P. (25 k and 40 k, Fig. 2a). The pollen record from Grande Pile, France, (Fig. 1) and Northern Hemisphere glacial evidence (Stuiver et al. 1978; Denton and Hughes 1981) likewise indicate climatic changes within this interval and at approximately 70,000 years ago. Whether these oscillations represent events similar in nature to the Younger Dryas is uncertain. To better understand the climate system and assess the possibility of future rapid climatic changes, we need to examine the distribution, mechanism, and possible cause(s) of the fluctuations.

The major cooling of the North Atlantic Ocean which took place resulted in the North Atlantic polar front advancing to the south and east, to a position about 5° poleward of its full glacial location (Ruddiman and McIntyre 1981). The GISS general circulation model (GCM) has been used in a series of experiments to test the impact of such colder North Atlantic surface ocean tem-

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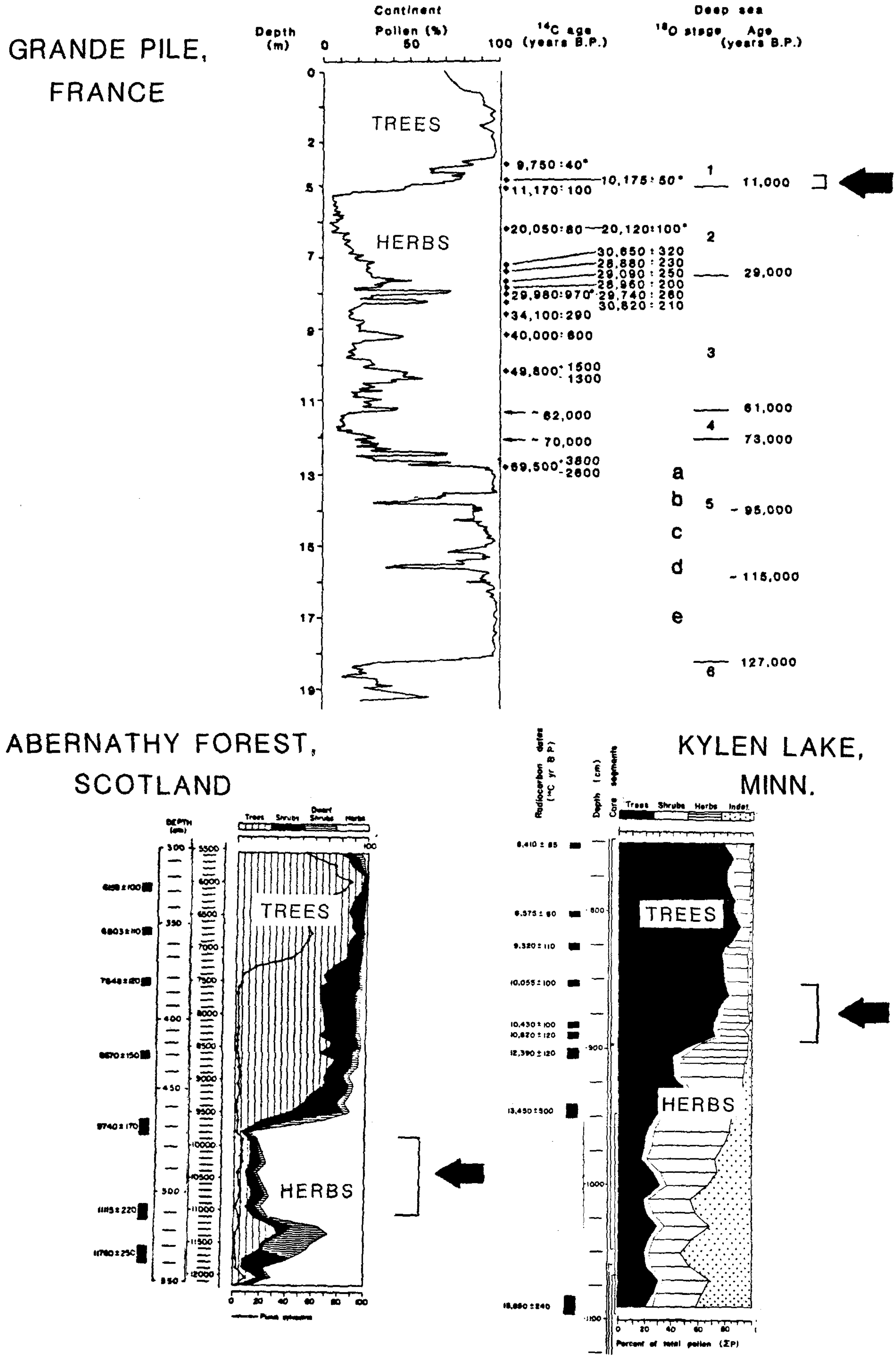


Fig. 1. Pollen percentage diagrams for Grande Pile, France (Woillard and Mook 1982) *(top)*, Abernathy Forest, Scotland (Birks and Mathewes 1978) *(left)*, and Kylen Lake, Minnesota (Birks 1981) *(right)*. The approximate Younger Dryas time interval, 11–10 k (11,000–10,000 B.P.), is bracketed and designated by an *arrow*. The Allerod/Younger Dryas tree/herb oscillation is present in the records from France and Scotland, but is not apparent in Minnesota

peratures on global climate. Several problems are addressed: What would the effect of colder ocean temperatures in this limited region have been on climate worldwide? How does this result compare with evidence of climate change suggested by pollen, glacial, and ice core records correlative with the Alleröd/Younger Dryas event? What is the magnitude of the temperature change both annually and seasonally between the Younger Dryas and today? What might be expected if the North Atlantic cooled in a similar manner in the future? Finally, what is needed to enhance our understanding of the Younger Dryas and other possible rapid climatic changes?

We first present a brief summary of terrestrial and oceanic evidence indicating the climatic changes which occurred around 11 k. We then describe the model, the experiments and the results. A comparison of terrestrial paleoclimatic estimates and model-computed temperature changes is then used to discuss questions concerning the nature and extent of the Younger Dryas cooling.

Terrestrial evidence for the Younger Dryas cooling (11–10 k)

Recognition of the Younger Dryas interval is stratigraphically often problematic for three major reasons: sampling interval selection, dating, and climatic interpretation of pollen fluctuations. Adequate sedimentation rates and detailed sampling are necessary to decipher rapid changes in lithology and vegetation (Watts 1983). Therefore, demonstration of the Younger Dryas interval in specific locations is often qualitative and subject to uncertainty. The following compilation represents available data concerning this interval, both in Europe where it is best recognized, and globally.

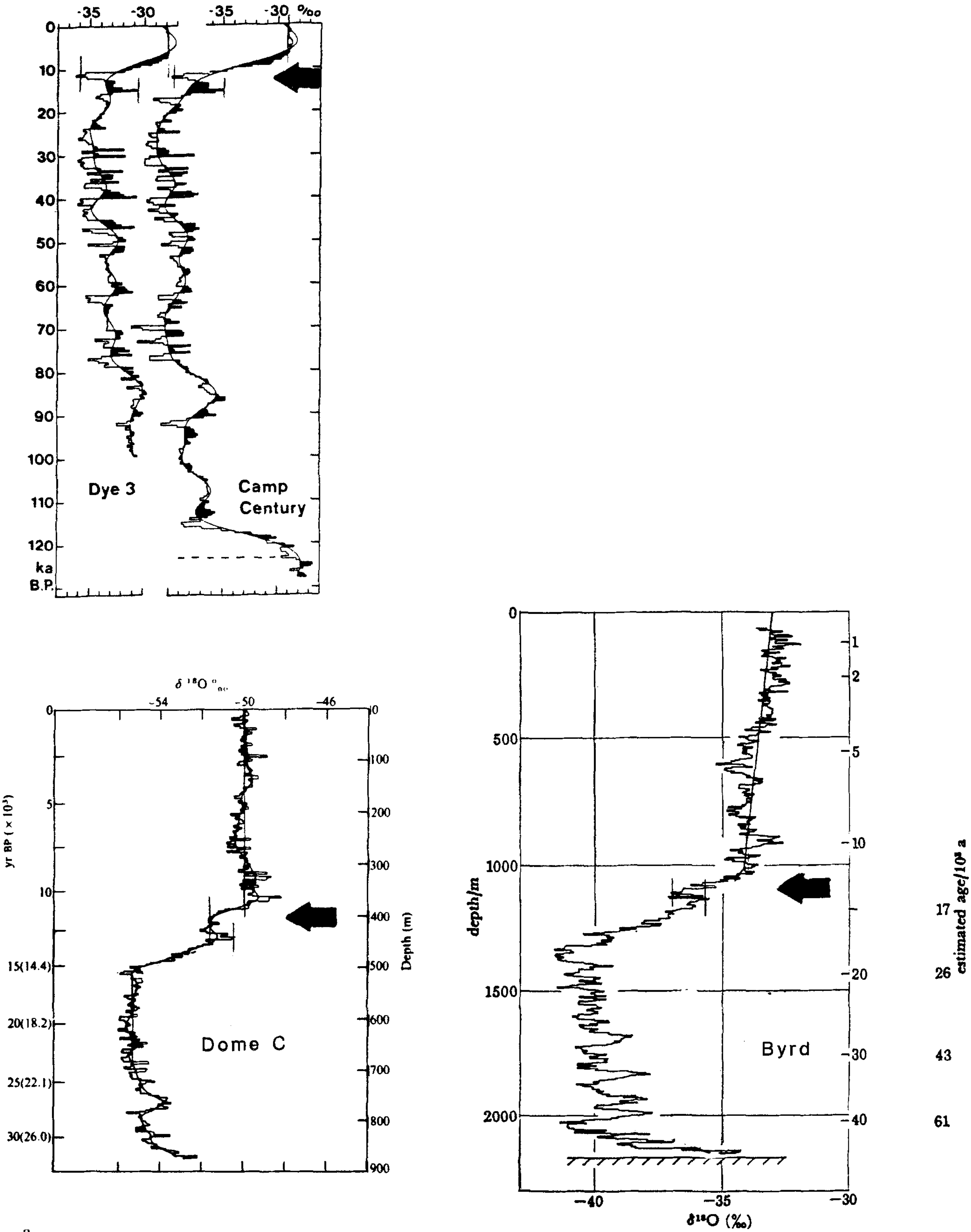
Estimates of climate change in Europe during the Younger Dryas

Evidence for late-glacial climatic fluctuations in western Europe has been documented and substantiated by numerous studies since the turn of the century (beginning with Hartz and Milthers 1901, as noted by Mercer 1969). Extensive geomorphological (i.e., glacial, permafrost, frost breakup of rocks, downslope erosion), faunal (i.e., beetles, crustaceans, cladocerans, vertebrates), floral (pollen, macrofossils), and isotopic (lake carbonates) evidence indicates a reversion to stadial conditions during the Younger Dryas which was most pronounced in the British Isles (Sissons 1979b; Watts 1980a). Warming approximately 13,000 B.P. was followed by a readvance of the Scandinavian ice sheet (Mangerud 1970) and regeneration of the Scottish Highlands glaciers (Sissons 1979a). Faustova (1983) documents the late-glacial eastern extension of the Scandinavian ice in Karelia and the Kola Peninsula, USSR. The northern European Younger Dryas ice front positions as summarized from several sources are shown by Larson et al. (1984 Fig. 9).

European pollen sites recording the Younger Dryas fluctuation within approximately 11,000–10,000 B.P., along with those indicating a questionable fluctuation, are illustrated in Fig. 3 (Table 1). Over 60 late-glacial pollen records are found in Scotland alone (Walker 1984), though many of these lack dates. Available estimates of mean July temperature depression are given in Table 4. Watts (1980a) summarizes the regional variation in the intensity of vegetational responses throughout Europe, emphasizing the major change in species composition that took place in the British Isles (Scottish pollen diagram in Fig. 1) i.e., the replacement of birch forest with shrubs and herbs. In contrast, many continental European records indicate relatively small changes in species abundances but a marked lithologic change from organic (Alleröd) to inorganic deposition, particularly in the Alps.

Further south in Europe, the record of a possible Younger Dryas cooling is questionable. Van der Hammen et al. (1971) contend that a northeastern Macedonian pollen diagram (Wijmstra 1969) shows a late-glacial climatic oscillation, though ^{14}C dates indicate the oscillation may have occurred during the Holocene (Van Zeist and Bottema 1982). Throughout the British Isles and the rest of Europe, numerous radiocarbon dates are synchronous with the Younger Dryas, but others are not. Chronological uncertainties in Scotland (Price 1983) indicate the Younger Dryas inception between 11,300–10,600 B.P. and its close anywhere between 10,250–9,750 B.P. The advent of the mass spectrometry accelerator dating technique should reduce many of these chronological uncertainties because individual seeds and needles may be dated.

Most of the descriptions of climate during the Younger Dryas relate to temperature effects. However, detailed evaluation of precipitation change is available for a few regions. From the altitude of firn line glaciers in the British Isles, Sissons (1979a) suggests that during the Younger Dryas the zone of maximum snowfall moved southward over the British Isles resulting in a deficiency in precipitation in the northern mainland of Scotland. From the distribution of cirque moraines and equilibrium-line altitudes, Larson et al. (1984) suggest that snow-bearing winds came from the southwest. They infer that an open North Atlantic and southern Norwegian Sea were the main precipitation sources for the Younger Dryas glaciers, and heaviest precipitation in Norway was in the Bergen-Nordford region with increased continentality both to the north and south.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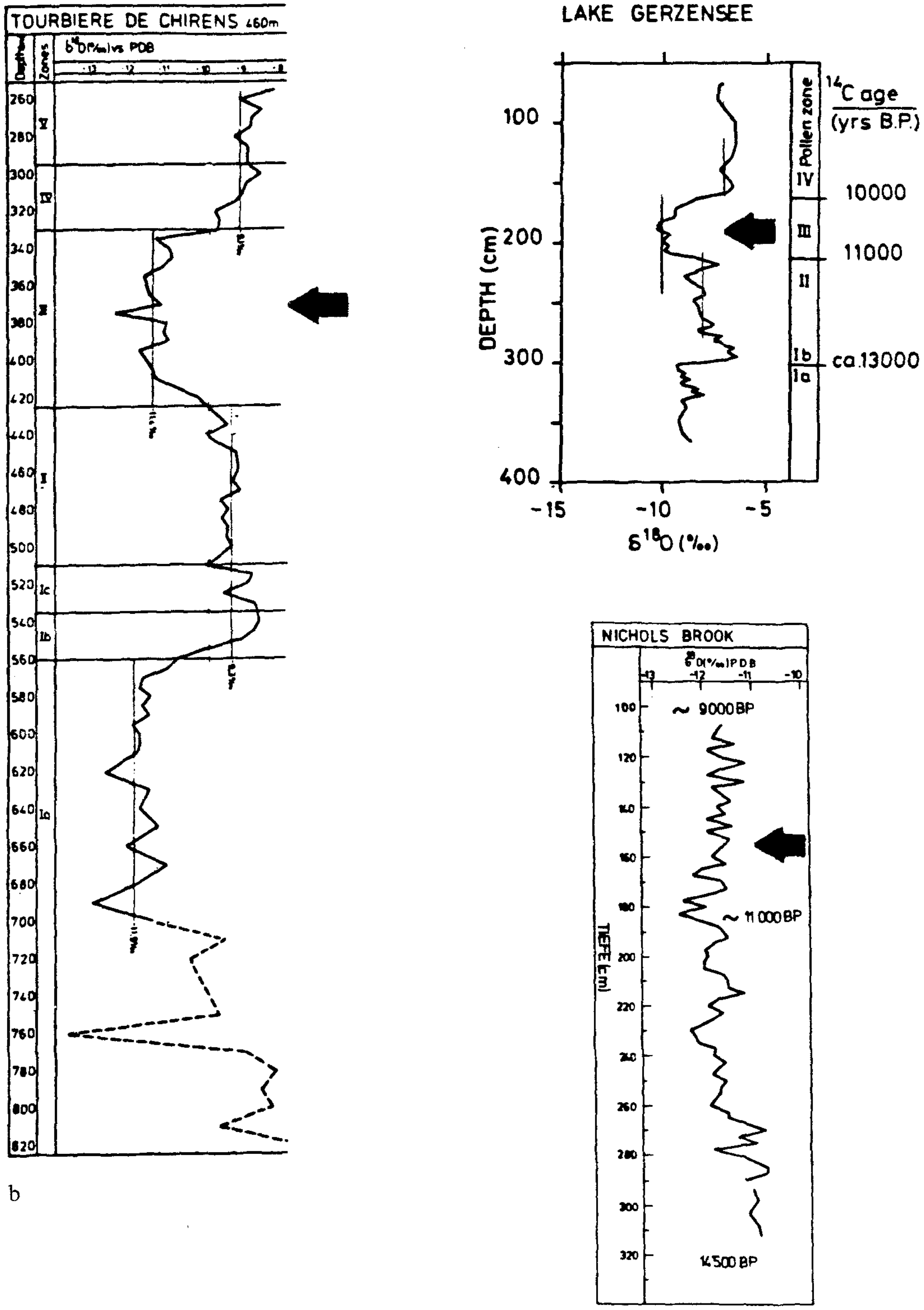


b

Fig. 2a. $^{18}O/^{16}O$ records from the Greenland *(top)* and Antarctic *(bottom)* ice cores. *Arrows* indicate a possibile Younger Dryas fluctuation, although dating is somewhat uncertain. The magnitude of the temperature change implied by the fluctuations can be calculated by assuming a 0.7‰ change per °C; **b** Oxygen isotope profiles in marl sediments of Tourbiere de Chirens, France (Eicher et al. 1981), Lake Gerzensee, Switzerland (Eicher 1980), and Nichols Brook, N.Y. (Eicher and Siegenthaler 1982). *Arrows* indicate an 11–10 k Younger Dryas fluctuation in Europe, but it is apparently absent at Nichols Brook, N.Y. The temperature change can be calculated by assuming a 0.5‰ change per °C

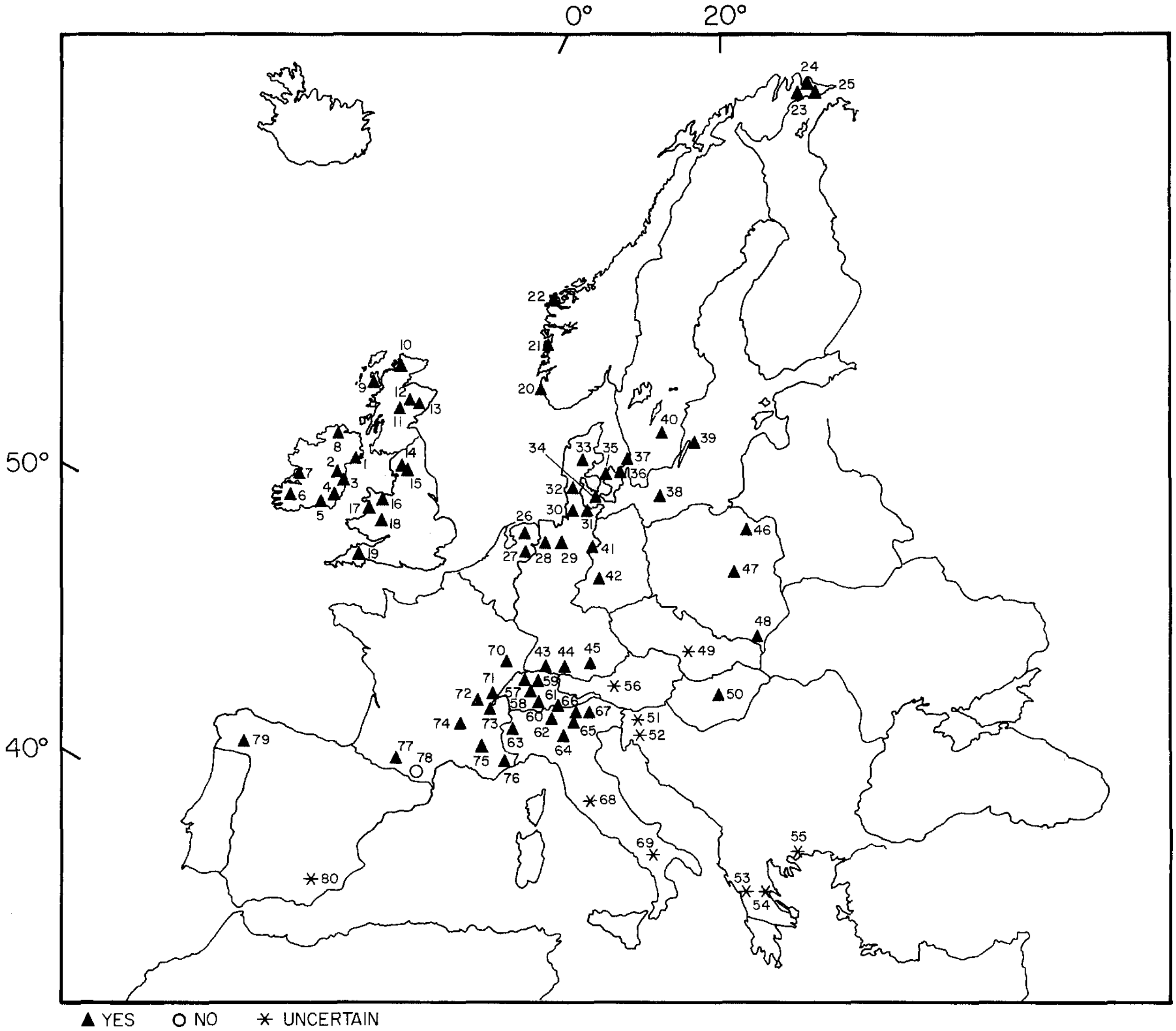


Fig 3. European fossil pollen sites indicate the presence *(closed triangle)*, absence *(open circle)* or possibility *(asterisk)* of a late-glacial Alleröd/Younger Dryas oscillation. Numbered sites refer to the listing in Table 1

Table 1. European palynological/lake-level climatic oscillations ca. 11 k

No.	Location	Oscillation	Reference
1	Woodgrange, Ireland	yes	Singh 1970
2	Dunshaughlin, Ireland	yes	Watts 1977
3	Ballybetagh, Ireland	yes	Watts, 1977
4	Coolteen, Ireland	yes	Craig 1978
5	Belle Lake, Ireland	yes	Craig 1978
6	Muckross, Ireland	yes	Watts 1977
7	Lough Goller, Ireland	yes	Watts 1977
8	Glenveagh, Ireland	yes	Watts 1977
9	Isle of Skye, Scot.	yes	Birks HJB 1973
10	Lochan an Druim, Scot.	yes	Birks HH 1984
11	Loch Etteridge, Scot.	yes	Birks and Mathewes 1978
12	Abernathy Forest, Scot,	yes	Birks and Mathewes 1978
13	Loch a 'Chnuich, Scot.	yes	Birks and Mathewes 1978
14	Low Wray Bay, Britain	yes	Pennington 1975, 1977
15	Blelham Bog, Britain	yes	Pennington 1977
16	Nant Ffracon, N. Wales	yes	Seddon 1962
17	Glanllyrnau, N. Wales	yes	Simpkins 1974

Table 1 (continued). European palynological/lake-level climatic oscillations ca. 11 k

No.	Location	Oscillation	Reference
18	Elan Valley, Wales	yes	Moore 1970
19	Bodmin Moor, Britain	yes	Brown 1977
20	Brondymyra, Norway	yes	Chanda 1965
21	Blomoy, Norway	yes	Mangerud 1970
22	Krakenes, Norway	yes	Larson et al. 1984
23	Bergebyvatnet, Norway	yes	Prentice 1982
24	Hollmfjellvatnet, Nor.	yes	Prentice 1982
25	Sternevatnet, Norway	yes	Prentice 1982
26	Waskemeer, Netherlands	yes	Casparie and Van Zeist 1960
27	Usselo, Netherlands	yes	van der Hammen 1951; Tauber 1960
28	Westrehauderfehn, Ger.	yes	Behre 1967
29	Elmshorn, Germany	yes	Usinger 1981
30	Glusing, Germany	yes	Menke 1968
31	Kubitzbergmoor, Ger.	yes	Usinger 1975
32	Rabensbergmoor, Ger.	yes	Usinger 1975
33	Bollingso, Denmark	yes	Iversen 1954
34	Fladet, Denmark	yes	Fredskeld 1975
35	Ruds Vedby, Denmark	yes	Krog 1954
36	Alleröd, Denmark	yes	Iversen 1954
37	Hakulls Mosse, Sweden	yes	Berglund 1971; 1976
38	Bornholm, Denmark	yes	Iversen 1954
39	Karslo Basin, W. Balt.	yes	Alm 1982
40	Akerhultagol, Sweden	yes	Berglund 1976
41	Siemen, Germany	yes	Lesemann 1969
42	Gaterlesbener See, Ger.	yes	Firbas et al 1955
43	Buchensee, Germany	yes (weak)	Bertsch 1961
44	Dietenberger See, Ger.	yes	Bertsch 1961
45	Kirchseeoner Moor, Ger.	yes	Rausch 1975
46	Mikolajki, Poland	yes	Ralska-Jasiewiczowa 1972
47	Witow, Poland	yes	Wasylikow 1964
48	Tarnawa, Poland	yes	Ralska- Jasiewiczowa 1972
49	Vracow, Czech.	?	Rybnickova and Rybnicek 1972
50	Dunakeszi, Hungary	yes	Jarai-Komlodi 1970
51	Kostanjevica, Yugo.	no	Sercelj 1971
52	Trstenik, Yugo.	no	Sercelj 1971
53	Ioannini, Greece	?	Bottema 1967
54	Tenaghi Phillipon, Greece	?	Wijmstra 1969
55	Xinias, Greece	?	Bottema 1978
56	Carinthia, Austria	yes	Fritz 1972; Schultze 1982
57	Gerzensee, Switzerland	yes (weak)	Eicher and Siegenthaler 1976
58	Obergurbs, Switz.	yes (weak)	Kuttel 1974
59	Faulenseemoos, Switz.	yes (weak)	Eicher and Siegenthaler 1976
60	Wachseldorn, Switz.	yes	Oeschger et al 1980
61	Campra, Switzerland	yes	Muller 1972
62	Bedrina, Switz.	yes	Kuttel 1977
63	Tourbiera di Trena	yes	Schneider 1978
64	Saltarino Sotto, Italy	yes	Gruger 1968
65	Lago di Ledro, Italy	yes	Beug 1964
66	Fiave, Italy	yes	Gruger 1968
67	Bondone, Italy	yes	Gruger 1968
68	Lago de Vico, Italy	?	Frank 1969
69	Monticchio, Italy	?	Watts, 1985
70	Grand Chemin, France	yes	Woillard 1975, 1979
71	Les Cruilles, France	yes	Wegmuller 1966
72	Lac de Chalain, France	yes	Wegmuller 1966
73	La Pile, France	yes	Wegmuller 1966
74	Massif Central, France	yes	de Beaulieu et al. 1982
75	Tourbiere de Chirens, Fr.	yes	Eicher et al. 1981
76	Lac Long, Inferieur, Fr.	yes	de Beaulieau 1977
77	Biscaye, France	yes	Mardones and Jalut 1983
78	Ariege, France	no	Jalut et al 1982
79	Lago de Sanabria, Spain	yes	Watts 1983; Hannon 1984
80	Padul, Spain	?	Florschutz et al. 1971

The Younger Dryas outside of Europe

While there is evidence for worldwide late-glacial warming (approximately 16,000–13.000 B.P.) from a variety of paleoclimatic indicators ($\delta^{18}O$, foraminifera, mollusks, glaciers, beetles, pollen), the concept of the Younger Dryas cooling as a worldwide event remains controversial (Moore 1981). Mercer (1969) and Wright (1977) contended that it was restricted to Europe. The same problems of sampling, chronology, and palynological interpretation make correlations difficult. Claims for the oscillation further eastward in the USSR are made by Faustova (1983) on the basis of glacial features and by Khotinskiy (1983) from Soviet pollen records. The easternmost site which hints of a climatic reversal (Moore 1981) is from Beijing, China (Kong and Du 1980), where a reversal in the trend toward a warmer climate prior to 10,750 B.P. may be correlative with the clasical European sequence.

Records in North America that encompass the Younger Dryas interval are shown in Fig. 4 (Table 2) and only those adjacent to the North Atlantic report unequivocal evidence of a climatic oscillation. Mott et al. (1984), Mott (1985) and Anderson (1983) report 20 sites from eastern maritime Canada with stratigraphic, pollen, and/or macrofossil evidence and radiocarbon control that suggest that this oscillation was felt in extreme eastern North America. They note the significance of a major readvance of Newfoundland ice (Grant 1969) at 10,900 B.P. Livingstone and Livingstone's (1958) early record from Nova Scotia also shows pollen fluctuations which can be interpreted as indicating a climatic oscillation. Further south, in New England, several pollen diagrams (nos. 33-42, Fig. 4) exhibiting sharp species oscillations approximately 11–10 k are suggestive of climatic change (Peteet, 1985) although no climatic reversals are reported. In the mid United States region, the southern states, and westward, no cooling effect appears in the vegetational response. One possible exception is the ecotonal region of glaciated Ohio, where Shane (1980) correlates a climatic shift from several pollen sites with the Alleröd-Younger Dryas.

On the western coast of North America from Vancouver Island south to Washington, several sites with marked species fluctuations (Heusser 1973; Mathewes 1973; Heusser 1974; Barnosky 1981; Hebda 1983) suggest a positive climatic correlation in either temperature or precipitation with the Younger Dryas elsewhere although only one (Hebda 1983) has good chronology. The glacial advance of the Sumas Stade in Washington, approximately

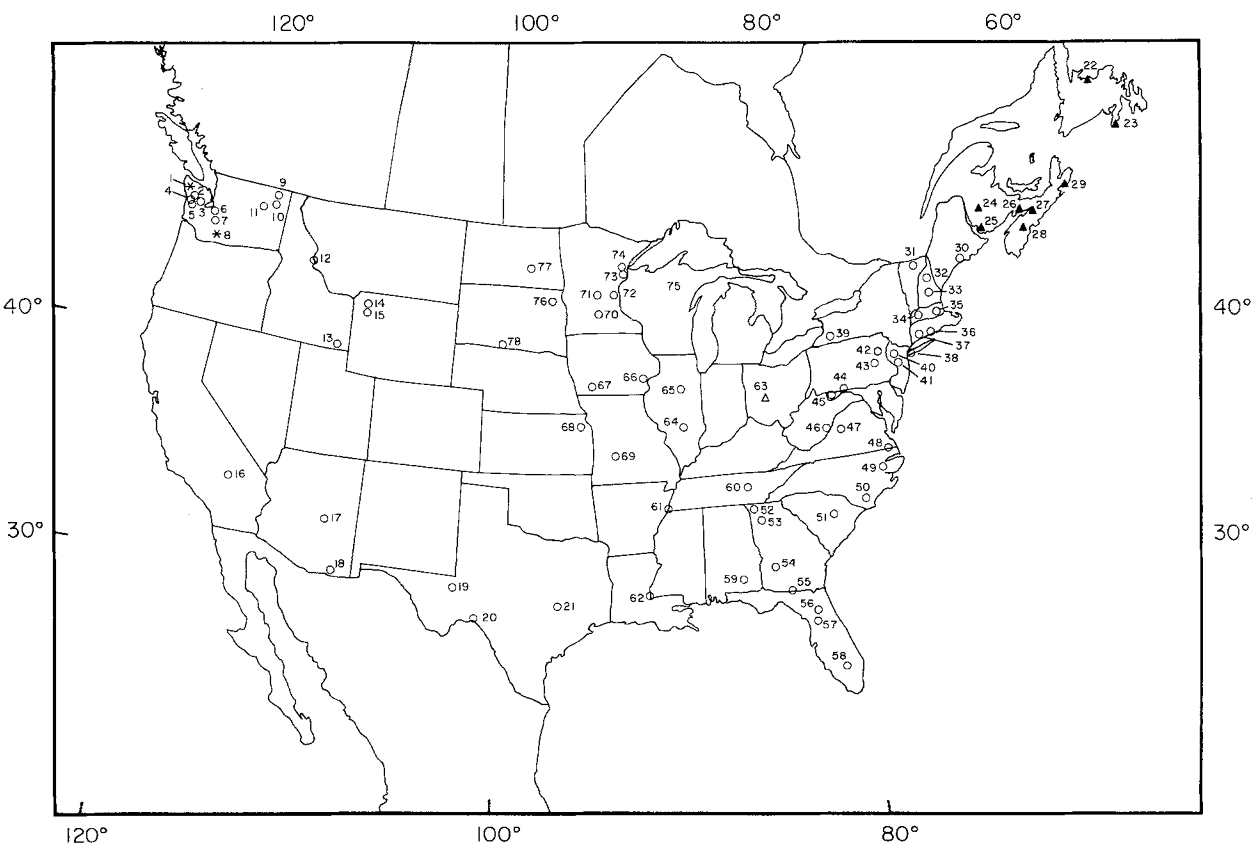


Fig. 4. North American fossil pollen sites indicating the presence (*closed triangle*), absence (*open circle*) or possibility (*asterisk*) of a late-glacial Alleröd/Younger Dryas oscillation. Numbered sites are listed in Table 2

Table 2. North American palynological climatic oscillations ca. 11 k

No.	Location	Oscillation	Reference
1	Wesseler, Wash.	?	Heusser 1973
2	Bogachiel, Wash.	no	Heusser 1983
3	Hoh, Wash.	no	Heusser 1972
4	Quinault, Wash.	no	Heusser 1983
5	Humptulips, Wash.	no	Heusser 1983
6	Nisqually Lake, Wash.	no	Heusser 1983
7	Mineral Lake, Wash.	no	Hibbert 1979
8	Davis Lake, Wash.	no	Barnosky 1981
9	Mud Lake, Wash.	no	Mack et al 1979
10	Waits Lake, Wash.	no	Mack et al 1978
11	Creston Fen, Wash,	no	Mack et al 1976
12	Lost Trail Bog., Wash.	no	Mehringer et al 1977
13	Swan Lake, Idaho	no	Bright 1966
14	Cub Creek Pond, Wyo.	no	Waddington and Wright 1974
15	Buckbean Fen, Wyo.	no	Baker 1976
16	Searles Lake, Calif.	no	Leopold 1967
17	Potato Lake, Ariz.	no	Whiteside 1965
18	Lehner Ranch, Ariz.	no	Mehringer and Haynes 1965
19	Crane Lake, Texas	no	Hafsten 1961
20	Bonfire Shelter, Texas	no	Bryant 1978
21	Boriack Bog, Texas	no	Bryant 1977; Bryant and Shafer 1977
22	Leading Tickles, Newf.	yes	Anderson 1983; Blake 1983; Anderson and Macpherson 1984
23	Burin Peninsula, Newf.	yes	Anderson 1983; Blake 1983; Anderson and Macpherson 1984
24	Roulston Lake, N. Bruns.	yes	Mott 1985
25	Basswood Rd. Lake, N. Bruns.	yes	Mott 1975, Mott 1985
26	Leak Lake, Nova Scotia	yes	Mott 1985
27	Brookfield Lake, N. Scotia	yes	Mott 1985
28	Canoran Lake, N. Scotia	yes	Mott 1985
29	Gillis Lake, N. Scotia	yes	Livingstone and Livingstone 1958
30	Moulton Pond, Maine	no	Davis et al. 1975
31	Brownington Pond, Vt.	no	Davis 1965
32	Mirror Lake, N. H.	no	Likens and Davis 1975; Davis and Ford 1981
33	Sandogardy Pond, N. H.	no	Davis 1983
34	Pleasant St. Bog, Mass. Tom Swamp, Mass. Gould's Bog, Mass.	no	Davis 1958
35	Taunton, Mass.	no	Davis 1960
36	Linsley Pond, Conn.	?	Deevey 1939
37	Rogers Lake, Conn.	no	Davis 1965, 1969, 1983
38	Kings Point Bog. N. Y.	no	Sirkin 1967
39	Belmont Bog, N. Y.	no	Spear and Miller, 1976
40	Francis Lake, N. Y.	no	Cotter 1983
41	Lawyers Bog, N. J.	no	Cotter 1983
42	Tannersville, Pa.	no	Watts 1979
43	Longswamp, Pa.	no	Watts 1979
44	Crider's Pond, Pa.	no	Watts 1979
45	Buckles Bog, Md.	no	Maxwell and Davis 1972
46	Cranberry Glades, W. Va.	no	Watts 1979
47	Hack Pond, Va.	no	Craig 1969
48	Dismal Swamp, Va.	no	Whitehead and Oaks 1979
49	Rockyhock Bay, N. C.	no	Whitehead 1973
50	Singletary Lake, N. C.	no	Whitehead 1967
51	White Pond, S. C.	no	Watts 1980a
52	Pigeon Marsh, Ga.	no	Watts 1975
53	Quicksand, Bob Black, and Green Ponds, Ga.	no	Watts 1970
54	Pennington, Ga.	no	Watts 1980b
55	Lake Louise, Ga.	no	Watts 1971
56	Sheelar Lake, Fla.	no	Watts and Stuiver 1980
57	Mud Lake, Fla.	no	Watts 1969
58	Lake Annie and Buck Lake, Fla.	no	Watts 1975
59	Goshen Springs, Ala.	no	Delcourt 1980
60	Anderson Pond, Tenn.	no	Delcourt 1979

Table 2 (continued) North American palynological climatic oscillations ca. 11 k

No.	Location	Oscillation	Reference
61	Nonconnah Creek, Tenn.	no	Delcourt et al 1980
62	Tunica Hills, La.	no	Delcourt and Delcourt 1977
63	Ohio	yes	Shane 1980
64	Pittsburg Basin and Semming School, Ill.	no	Gruger 1972a; 1972b
65	Richland Creek, Ill.	no	Gruger 1972b
66	Butler Farm, Iowa	no	Van Zant et al 1980
67	Brayton, Iowa	no	Baker et al 1980
68	Arrington and Muscotah Marsh, Kansas	no	Gruger 1973
69	Boney Spring, Mo.	no	King 1973
70	Norwood, Minn.	no	Ashworth et al 1981
71	Wolf Creek, Minn.	no	Birks 1976
72	White Lily Lake, Minn.	no	Cushing 1967; Birks 1976
73	Lake Kotiranta, Minn.	no	Wright and Watts 1969
74	Kylen Lake, Minn.	no	Birks 1981
75	Wood Lake, Wisc.	no	Heide 1984
76	Pickeral Lake, S. D.	no	Watts and Bright 1968
77	Tappen, N. D.	no	Wright 1970
78	Rosebud, S. D.	no	Watts and Wright 1966

11,500–10,000 B.P. (Armstrong 1981) may support an interpretation of regional cooling in the Pacific Northwest (Hansen and Easterbrook 1974; Heusser 1977).

S. Georgia in the subantarctic was subject to a late-glacial fluctation in ice extent prior to 9,500 B.P. (Clapperton et al. 1978) and a similar fluctuation is suggested by Clapperton and Sugden (1982) for West Antarctica, although dates are missing. Of major interest also are the Dome C, Byrd, and Vostok ice cores from Antarctica, which all exhibit possible late-glacial oscillations in $^{18}O/^{16}O$ (Fig. 2a), though lack of an absolute chronology leaves uncertainties about their synchroneity with the Greenland ice cores and the European sequence. The amplitude of the oscillation in the Antarctic ice cores is muted compared with that in the Greenland ice cores, as is the glacial-interglacial change. The Antarctic cores are from locations further removed from warmer ocean waters than are those from Greenland and are thus presumably less subject to cooling-induced isotopic moisture changes. Note (Fig. 5) that while the evidence is inconclusive, there are some indications of late-glacial climatic fluctuations south of 30°S in a variety of locations.

Palynological and glacial evidence for a worldwide Younger Dryas cooling event is not convincing. Based upon a cursory review of the 13–10 k data, locations are shown (Fig. 5; Table 3) where an oscillation in either temperature or moisture is recorded, absent, or considered a possibility. In many of these sites (Table 3) various problems such as sampling selection, C^{14} resolution (e.g., Lowe and Walker 1980), or topographic setting preclude identification of the fluctuation as correlative with the Alleröd/Younger Dryas. These compilations

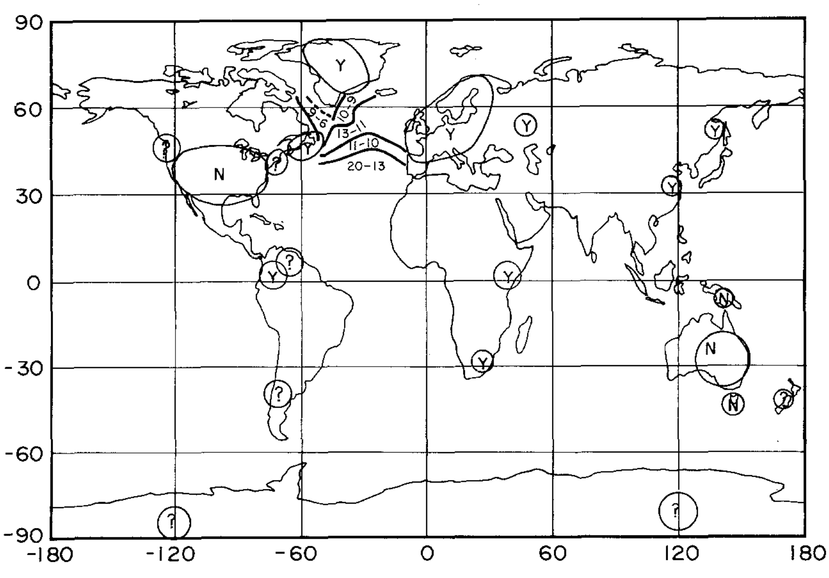


Fig. 5. Worldwide assessment of the published results indicating the presence (Y), absence (N), or possible (?) paleoclimatic indication of a late-glacial climatic oscillation which may or may not be correlative with the Alleröd-Younger Dryas. Blank areas indicate regions where an assessment has not been reported. Also included is the location of the North Atlantic polar front for different time periods (Ruddiman and McIntyre 1981). Deglacial retreat was interrupted by the readvance from 11–10 k

Table 3. Paleoclimatic evidence for late-glacial climatic oscillation outside Europe and North America

Location	Evidence	Oscillation	Reference
USSR			
Polovesko-Kupanskoye	pollen	yes	Khotinskiy 1983
Tula Bog	pollen	possible	Khotinskiy 1983
Uandi Mire (east USSR)	pollen	possible	Khotinskiy 1983
Kola Peninsula	glacial	yes	Faustova 1983
CHINA			
Beijing	pollen	possible	Kong and Du 1980, Moore 1981
JAPAN			
Central Japan	pollen, diatoms	yes	Fuji 1982
AFRICA			
Lake Chad	lake level	yes (high, low, high)	Kendall 1969
Lake Victoria	lake level	yes (high, low, high)	Maley 1977
Lake Kivu	lake level	yes (high, low, high)	Gillespie et al. 1983
Ethiopian Rift Lakes	lake level	yes (high, low, high)	Street-Perrott and Roberts 1983; Gillespie et al. 1983
N. Botswana	geomorphology	yes	Cooke 1984
Mt. Kenya	pollen	yes (wet, dry, wet)	Coetzee 1967
S. Africa	pollen	yes	Scott 1982; Butzer 1984
SOUTH AMERICA			
Peru, Colombia, Bolivia	glacial	yes	Clapperton 1983
Peru	glacial	yes	Wright 1984
Chile	glacial	no	Mercer 1984
Chile	pollen	yes	Heusser 1984
Chile	beetles	no	Hoganson and Ashworth 1982
Venezuela	pollen	possible	Bradbury et al. 1981
NEW GUINEA	glacial	no	Hope 1983
	pollen	no	Hope 1983
AUSTRALIA	pollen	no	Kershaw 1976; Kershaw et al. 1983
TASMANIA	pollen	no	Colhoun 1983
NEW ZEALAND	glacial	yes	Burrows 1979
	pollen	no	McGlone 1983
S. GEORGIA	glacial	yes	Clapperton et al. 1978

emphasize the need for a careful reassessment of the evidence. While scattered sites show marked fluctuations, the majority of pollen sites encompassing this time interval, particularly in North America, do not report such an oscillation. The possibilities of detecting ocean-atmosphere links, particularly between the North Atlantic, the Pacific, and the Antarctic, suggest that the search for additional sites with good chronological control would be worthwhile.

The magnitudes of temperature change reported for the different locations are given in Table 4. Uncertainties in the magnitudes are generally not available because of the difficulty in quantifying paleoclimatic estimates of temperature and precipitation from palynological records. Moore (1980) and Watts (1980a, 1983) discuss several of the reasons for this difficulty with particular regard to pollen records in Europe, i.e., identification of taxa only to the generic level, a lack of utilization of macrofossils to augment the pollen record, and lack of modern analogs for reconstruction of the late-glacial sequences. These problems also thwart the use of transfer functions (e.g., Webb and Bryson 1972) or other methods to reconstruct past temperature or precipitation. Hence, critical indicator species are used, assuming constancy in ecological tolerances over time. Ecotonal areas generally show greatest sensitivity to climatic change, but migrational and successional factors often complicate the interplay of species occurrence and percentage variation. Table 4 indicates that more studies demonstrate colder conditions than changes in precipitation.

Oceanographic evidence for North Atlantic cooling

Late-glacial oscillations in planktonic foraminifera (*Globigerina pachyderma*) from North Atlantic cores have been interpreted to indicate a large sea surface cooling associated with a southward readvance of the polar front during the Younger Dryas (Fig. 5; Ruddiman and McIn-

Table 4. Quantitative estimates of late-glacial paleoclimatic oscillations

Location	Reference	Evidence	Paleoclimate estimate (compared to present)		Model result 11kC,W–0kW
W. Norway	Larson et al. 1982	pollen and glacial	Y.D. summer temp	–4° to –5°C	–8°C
			winter precip.	–300mm	–150mm
	Larson et al. 1984	glacial	Y.D. summer temp	–6° to –8°C	–8°C
			if precip. = present amount		
Northern England	Coope 1977	beetles	ALLERÖD July temp	–3°C	0°C
SW Scotland			Y.D. July temp	–6°C	–6°C
Scotland	Sissons and	glacial	Y.D. July temp	–9°C	–8°C
SE Grampions	Sutherland 1976		if precip. = present amount		
Mull	Gray and Lowe 1977	glacial	Y.D. July temp	–10°C	–10°C
W Grampions	Ballentyne 1984	glacial	Y.D. Jan Temp	–27° to –17°C	–12°C
	Sissons 1979b, 1980	glacial	Y.D. July temp	–8°C	–8°C
Central England	Coope 1977	beetles	ALLERÖD July temp	–3°C	0°C
N Wales			Y.D. July temp	–6°C	–6°C
S England	Coope 1977	beetles	ALLERÖD July temp	–3°C	0°C
			Y.D. July temp	–6° to –8°C	–4°C
Ireland	Watts 1983	glacial	Y.D. annual temp	–7°C	–8°C
SE France	Eicher et al. 1981	pollen	Y.D. summer temp	–2° to –3°C	–1°C
S France	Jalut et al. 1982	glacial	ALLERÖD July temp	–4° to –6°C	+1°C
		pollen	Y.D. July temp	–4° to –5°C	–1°C
Poland	Mardones and Jalut 1983	pollen	ALLERÖD July temp	–3°C	~ –2°C
		soils	Y.D. July temp	–7°C	~ –2°C
		periglacial			
USSR	Khotinskiy 1983	pollen	Y.D. "short hot summers		+10°/+2°C
			cold, long winters"		–12°/–2°C
					west/east
Chile, central	Heusser and Streeter 1980	pollen	Y.D summer temp	–10°C	0°C
	Heusser 1982				
S. Chile	Heusser 1984a, b	pollen	Y.D. summer temp	–5° to –8°C	0°C
	Hoganson and Ashworth 1982	beetles	late-glacial temp	0°C	0°C
New Zealand	Burrows 1979	glacial	Y.D. annual temp	–2.5°C	0°C
	Burrows and Gellatly 1982				
Greenland					
Dye 3	Dansgaard et al. 1984	ice core	ALLERÖD annual temp	–4°C*	–4°C
			Y.D. annual temp	–11°C*	–7°C
Camp Century	Dansgaard et al. 1984	ice core	ALLERÖD annual temp	–8°C*	–2°C
			Y.D. annual temp	–14°C*	–3°C
Antarctica					
Byrd	G. de Q. Robin 1977	ice core	ALLERÖD annual temp	–3°C*	0°C
			Y.D. annual temp	–4°C*	0°C
Dome C	Lorius et al. 1979	ice core	ALLERÖD annual temp	–2°C*	0°C
			Y.D. annual temp	–3°C*	0°C

* We calculate temperature change estimates from the data in Fig. 2a using an average 0.7‰ change per °C. These estimates likely contain errors of at least a factor of two; note that the references cited do not quantify a Younger Dryas (Y.D.) temperature change

tyre 1973, 1981; Sancetta et al. 1973; Ruddiman et al. 1977; Duplessy et al. 1981). An earlier deglacial warming and polar-front retreat at 13,000 B.P. was followed after 11,500 B.P. by a cooling culminating between 11,000 and 10,000 B.P. Molluscan data confirm the existence of the polar front south of southwestern Norway at 10,300 B.P. (Mangerud 1977). The cause of this ocean surface cooling has not been definitely established (Ruddiman and McIntyre 1981). These authors as well as others (Mercer 1969; Grosswald 1980; Denton and Hughes 1981) suggested that the change may have resulted from a major influx of tabular icebergs associated with a disintegrating Arctic ice shelf or marine-based ice sheets, perhaps in the Barents Sea or the Kara Sea. Broecker et al. (1985) speculated that an influx of cold, fresh surface water could have also limited deep water formation, further accentuating the oceanic cooling – the water, upon losing heat to the atmosphere, would no longer sink and be replaced by warmer water from the south. Other explanations include the amplified expression of a 2,500-year climatic cycle suggested for montane Holocene glaciers (Denton and Karlen 1973), rearrangement of Rossby wave patterns and associated lower atmospheric circulation in the northern hemisphere due to melting ice sheets (Manley 1971), and free oscillations of the climate system (Ghil 1980). Regardless of the reason it has been widely assumed that the colder North Atlantic and the apparent climatic cooling over land were closely related (e.g., Sissons 1979; Watts 1980a; Mangerud 1980; etc.).

The actual sea surface temperatures for the 11 k time period have yet to be established in a systematic ocean wide fashion. For the purposes of the following experiments, we have returned the North Atlantic to its temperature distribution as determined by CLIMAP (1981) for the last glacial maximum. A comparison with spot estimates (Ruddiman and McIntyre 1981) shows that this may underestimate the Younger Dryâs temperatures somewhat.

Model and sensitivity experiments

The general circulation model used for these experiments is that described by Hansen et al. (1983). The model solves the equations for conservation of mass, energy, momentum, and moisture and calculates the radiative fluxes, cloud cover, surface fluxes, etc. It has realistic topography on an $8° \times 10°$ (latitude by longitude) horizontal resolution with fractional grid representation for land and ocean. Ground temperature calculations include the diurnal variation and seasonal heat storage, while ground hydrological parameters are a function of vegetation type. As shown in Hansen et al. (1983), the model produces realistic temperature fields when modern sea surface temperatures are used. A 5 year simulation of the modern climate represents the control run (0 kW, where W indicates that the present, relatively warm, temperatures were used in the North Atlantic) for the first experiment. As the results for the current climate have been shown elsewhere (e.g., Hansen et al. 1983), they are not reproduced here.

To determine the influence of colder North Atlantic ocean temperatures on climate, two sensitivity experiments were performed. In the first, (0 kC), we replaced the specified ocean temperatures for each month of the modern climate with Ice Age values derived from CLIMAP (1981) in the region of the North Atlantic north of 25°N. A sine curve fit to the CLIMAP August and February values was used to produce sea surface temperatures for the intervening months. While the polar front had not advanced quite as far south as it had during the last glacial maximum (Fig. 5), and temperatures were not quite as cold, this represents a first order approximation of the sea surface temperature field. Due to uncertainties in the sea ice field for this time, no change was made in the sea ice cover. Increasing sea ice would have further reduced latent and sensible heat fluxes and increased the surface albedo by some 30% in July. The other model boundary conditions were unaltered; sea level, land ice, orbital parameters, and sea surface temperatures in other regions were all kept at the modern values. This experiment thus isolates the effect of colder North Atlantic temperatures during modern conditions; however, it is important to note that by fixing the sea surface temperatures in other regions we are muting the potential climatic changes in areas away from the North Atlantic. The model was run for four years, and the results reported here are averages over the last three years.

Would the effects have been similar at 11 k? To answer this question a second set of sensitivity experiments was performed using specific 11-k boundary conditions. The 11-k orbital parameters (Berger 1978) and land ice distribution (Denton and Hughes 1981) along with a sea level change (amounting to about two-thirds of the drop in sea level which occurred in full glacial conditions) were incorporated into a control run (11 kW) with modern sea surface temperatures and into a run (11 kC) with the colder North Atlantic temperatures used in 0 kC. The difference between 11 kC and 11 kW represents to first order the effects of North Atlantic cooling during Younger Dryas-Alleröd conditions. The difference between 11 kC and 0 kW indicates the change in climate between the Younger Dryas and the present, with the important caveat that a more accurate Younger Dryas simulation must await the preparation of a complete data set for 11-k ocean surface temperatures as well as other boundary conditions (e.g., sea ice, vegetation, and perhaps CO_2). The difference between 0 kC and 0 kW provides an estimate of climate changes which might occur if the North Atlantic were to cool suddenly in the near future. Results for both experiment 11 kC and its control run represent 3-year averages. A description of the boundary conditions for each run is presented in Table 5.

Table 5

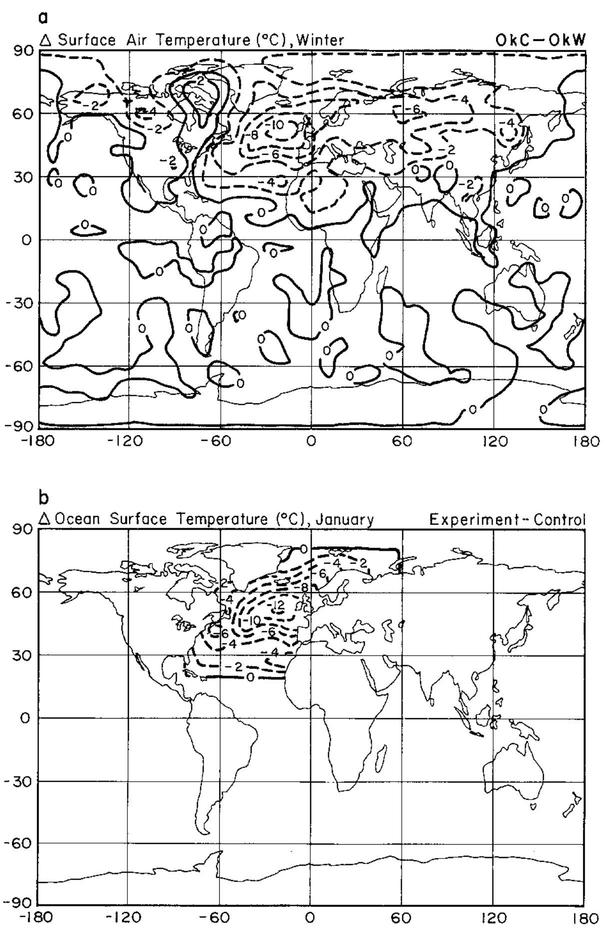
Designation	Description
0 kW	current climate (Hansen et al. 1983)
0 kC	current climate with 18-k sea surface temperatures for the North Atlantic north of 25° N (CLIMAP 1981)
11 kW	11-k orbital parameters (Berger 1977) 11-k land ice (Denton and Hughes 1981) current sea surface temperatures
11 kC	11-k orbital parameters, land ice; 18-k sea surface temperatures for the North Atlantic

Results and discussion

0 kC–0 kW

Surface air temperature changes resulting from cooler North Atlantic sea surface temperatures are shown in Figs. 6–8. The change in Northern Hemisphere winter (Fig. 6a) can be compared with the altered sea surface temperatures (Fig. 6b). The air temperature changes are largest over the North Atlantic (with a very similar pattern to the sea surface temperature changes), but are also very large downstream over central Europe. The temperature changes east of about 90°E are not significant (i.e., less than several standard deviations of the interannual changes found in the control run). The summer surface air temperature changes (Fig. 7a) are more strictly confined to the region of sea surface temperature change (Fig. 7b); the weaker zonal wind structure in summer produces less downstream advection than in winter, and solar radiation becomes important in influencing temperatures over land. There is thus less temperature change inland from the European coast in summer than in winter. The annual average surface air temperature change (Fig. 8a) can be compared with the standard deviations from the control run (Fig. 8b). Significant cooling (i.e., more than several standard deviations) occurs from approximately 30°–70° N and 70°W–50°E and further east at the higher latitudes. Note that little effect is seen over North America, except over extreme northeastern regions.

One direct result of the sea surface temperature changes is a relative stabilization of the atmosphere over the colder ocean waters – reducing the intensity of low pressure systems and increasing that of high pressure systems. The change in the annual average sea level pressure field (Fig. 9a) shows a positive increment over the North Atlantic and downstream, significant compared to the standard deviation (Fig. 9b). This result is reminiscent of the



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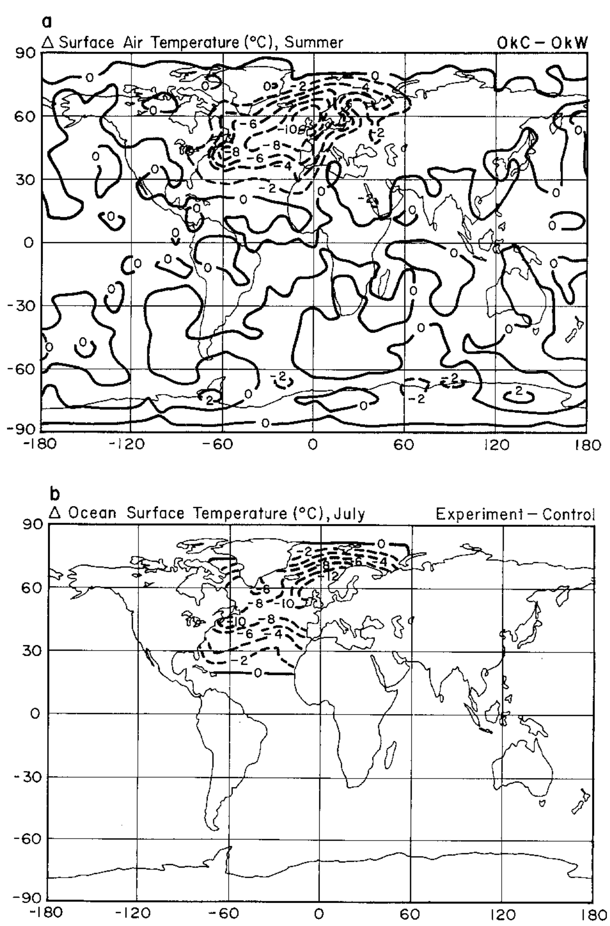

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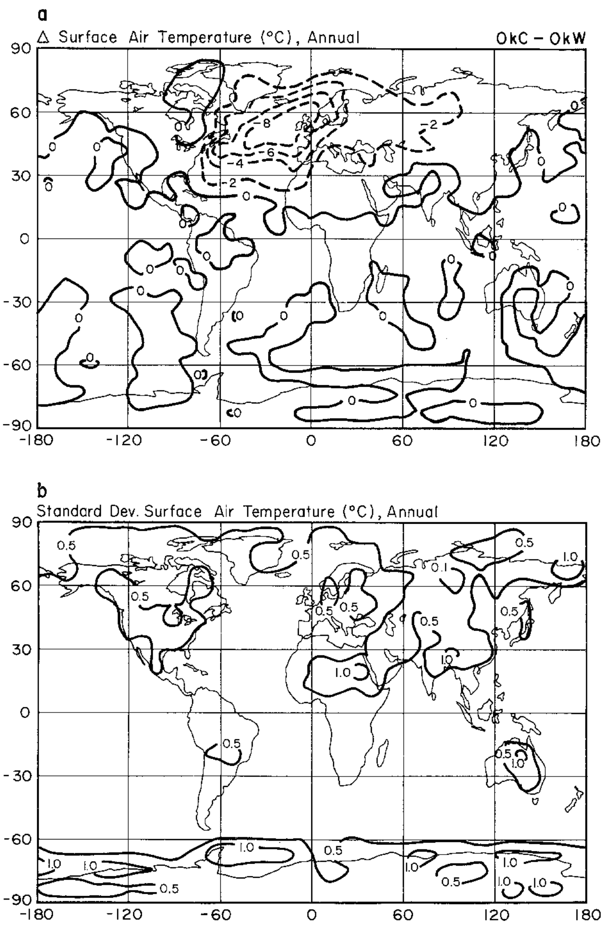
Fig. 6.a Surface air temperature change during winter (Dec–Feb), 0 kC–0 kW. Results are for three years averages. **b** Ocean surface temperature change for January, experiment minus control

Fig. 7.a Surface air temperature change during summer (June-Aug), 0 kC–0 kW. **b** Ocean surface temperature change for July, experiment minus control



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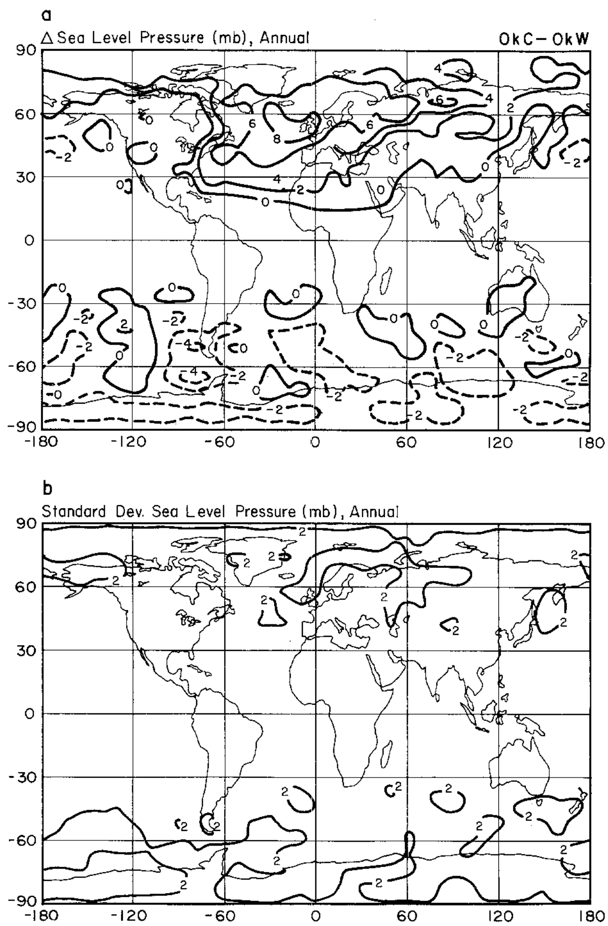

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Fig. 8.a Surface air temperature change, annual average, 0 kC–0 kW. **b** Standard deviation of annual average surface air temperature from five years of the control run

Fig. 9.a Sea level pressure change, annual average, 0 kC–0 kW. **b** Standard deviation of annual average sea level pressure

change shown by Kingston (1975) in the reconstructed weather map series for 1781–1786 during the Little Ice Age. The climate of that time apparently featured a high frequency of blocking anticyclones. Such a circulation would provide periods of uninterrupted northerly winds bringing polar air south, accentuating the cooling over western Europe. A similar effect can be seen somewhat in the model results (Fig. 10a). Due to the increased anticyclonic flow around high pressure systems in the western North Atlantic, the winds also become more easterly (onshore) over eastern North America, increasing the colder ocean influence (Fig. 10a). Thus, due to the influence of colder sea surface temperatures on atmospheric dynamics, adjacent landmasses experience enhanced cooling. Changes in regions far removed from the North Atlantic are, in general, not significant (standard deviations for the annual mean surface wind in the model vary from 0.1 ms^{-1} at low latitudes to 1 ms^{-1} in the South and North Pacific; in the North Atlantic they are generally around 0.4 ms^{-1}).

The insertion of cold sea surface temperatures creates a change in the low-level thermal contrast which affects all levels. The jet stream increases at the location of the increased latitudinal temperature gradient, displaying more of a west-to-east orientation across the Atlantic near 30°N (Fig. 10b). In this case some of the changes both downstream and upstream are of marginal significance, indicating that the altered North Atlantic temperatures may be affecting the upper air long-wave pattern with consequences for flow at various longitudes. Again, the changes in the Southern Hemisphere are not significant.

Associated with the increased sea level pressure is a decrease in the precipitation (Fig. 11a). The decrease north of 30°N over the Atlantic Ocean and western Europe is about five standard deviations (Fig. 11b) and thus highly significant, amounting to a reduction in precipitation of about 25%. The change over the ocean is greatest in winter, while over western Europe it is largest in summer. The precipitation increase to the south is a

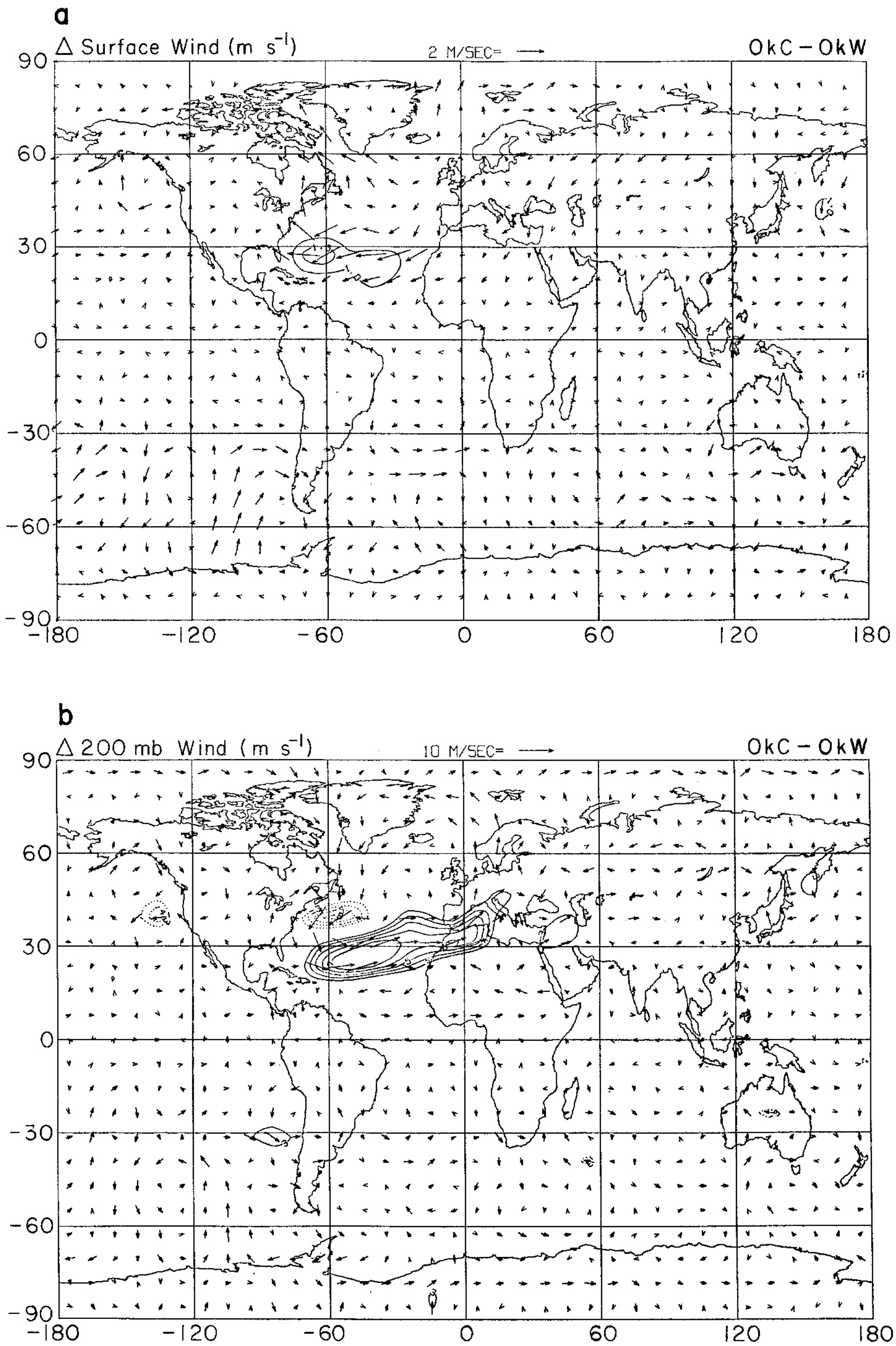


Fig. 10.a Surface wind change, annual average, 0 kC–0 kW. *Arrows* show change in wind direction, contours change in wind magnitude. **b** 200-mb (jet stream level) wind change, annual average, 0 kC–0 kW

consequence of the increased thermal contrast and change in jet stream, i.e., storms are directed by the jet stream over this location rather than moving further north. The colder ocean temperatures also lead to decreased evaporation, as discussed below. Despite reduced winter precipitation, northwestern Europe experiences an increase in winter snow depth of up to 50% due to the colder temperatures.

Shown in Table 6 are the annual changes for specific regions, along with the standard deviations from the control run (in parenthesis). As noted previously, the cooling is largest near the sea surface temperature anomaly, although the effect is still significant in the Siberian plateau. The mid United States and the North Pacific show little change. Significant precipitation decrease does not extend as far inland as temperature decrease. Despite the

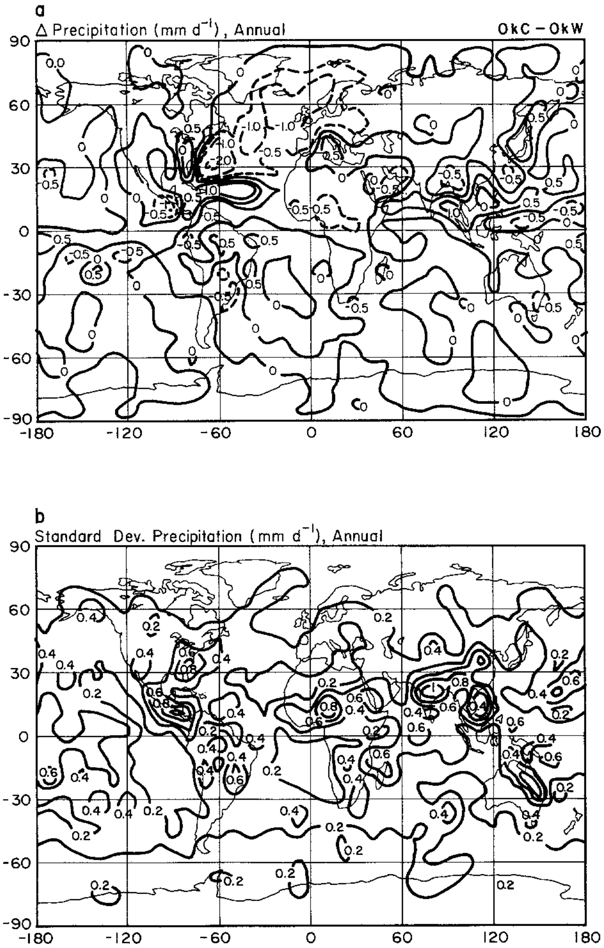


Fig. 11.a Precipitation change, annual average, 0 kC–0 kW. **b** Standard deviation of annual average precipitation

reduction in precipitation, the Atlantic Ocean experiences an increase in precipitation minus evaporation; were all else the same, it would have become fresher.

The remaining results relate to the change in energy balance. The increase in cloud cover over the North Atlantic diminishes the solar radiation absorbed at the surface. However, the large decrease in evaporative heat loss more than compensates for the change in radiation absorption, and the net heat flux into the ocean increases substantially. Because the ocean in this region normally loses heat to the atmosphere, the colder North Atlantic ocean waters are thus releasing less heat; if ocean transports remained fixed, the ocean in this region would have warmed.

This experiment provides a clear indication of the impact of colder ocean temperatures in the North Atlantic on the current climate system. Next we investigate whether similar changes would have occured at the Alleröd/Younger Dryas transition.

Table 6. Differences in annual average quantities and interannual standard deviations 0 kC–0 kW

	Middle US	North Atlantic	Middle Europe	Siberian plateau	North Pacific
Surface air temperature (T_S) (°C)	0.1 (0.1)	−5.9 (0.0)	−5.1 (0.3)	−1.9 (0.4)	0.0 (0.0)
Precipitation (P) (mm d^{-1})	0.0 (0.1)	−0.9 (0.1)	−0.6 (0.1)	−0.1 (0.1)	0.1 (0.1)
Precip-evap (P–E) (mm d^{-1})	0.0 (0.1)	0.4 (0.1)	−0.1 (0.2)	−0.1 (0.0)	0.0 (0.4)
Cloud cover (C) (%)	1.0 (1.4)	7.0 (1.2)	0.0 (0.9)	0.0 (1.1)	1.0 (0.4)
Net radiation at surface (W m^2)	−1.0 (1.0)	−4.0 (1.4)	1.0 (1.0)	0.0 (0.8)	0.0 (1.0)
Net heat at surface (W m^2)	0.0 (0.4)	37.0 (4.8)	5.0 (2.4)	0.0 (0.3)	−2.0 (4.8)

11 kC–11kW

Both of these model simulations incorporated the orbital parameters and land ice and sea level values appropriate for 11 k. The difference between them is the colder North Atlantic sea surface temperatures used in experiment 11 kC (e.g., Figs. 6b, 7b). To some extent, these conditions represent the boundary conditions that prevailed at the Alleröd/Younger Dryas transition with the caveats noted previously.

The changes in surface air temperature induced by the colder North Atlantic waters are shown in Fig. 12 (a–c) for Northern Hemisphere winter, summer, and the annual average. Comparison with the changes induced in the current climate (Figs. 6a, 7a, 8a) indicates that the colder sea surface temperatures produce very similar cooling in both climatic regimes. The magnitudes and patterns of change are almost identical. Once again the downstream influence is greater in winter, and again the cooling over North America is limited to the extreme northeastern sections. The location of the zero temperature change contour lies along eastern North America in both Figs. 12c and 8a, indicating the tenuous nature of the North American cooling result and emphasizing the importance of the proper specification of western North Atlantic sea surface temperature changes.

Changes in the annual average sea level pressure field, wind field, and precipitation (Figs. 13, 14a) are also very similar to the 0 kC–0 kW changes (Figs. 9, 11). The colder seawater temperatures lead to increased high pressure over the ocean, with reduced precipitation (and evaporation). The precipitation increase over the Atlantic just to the south of the colder waters follows the changed jet stream. As shown in Fig. 14b, the evaporation decrease is

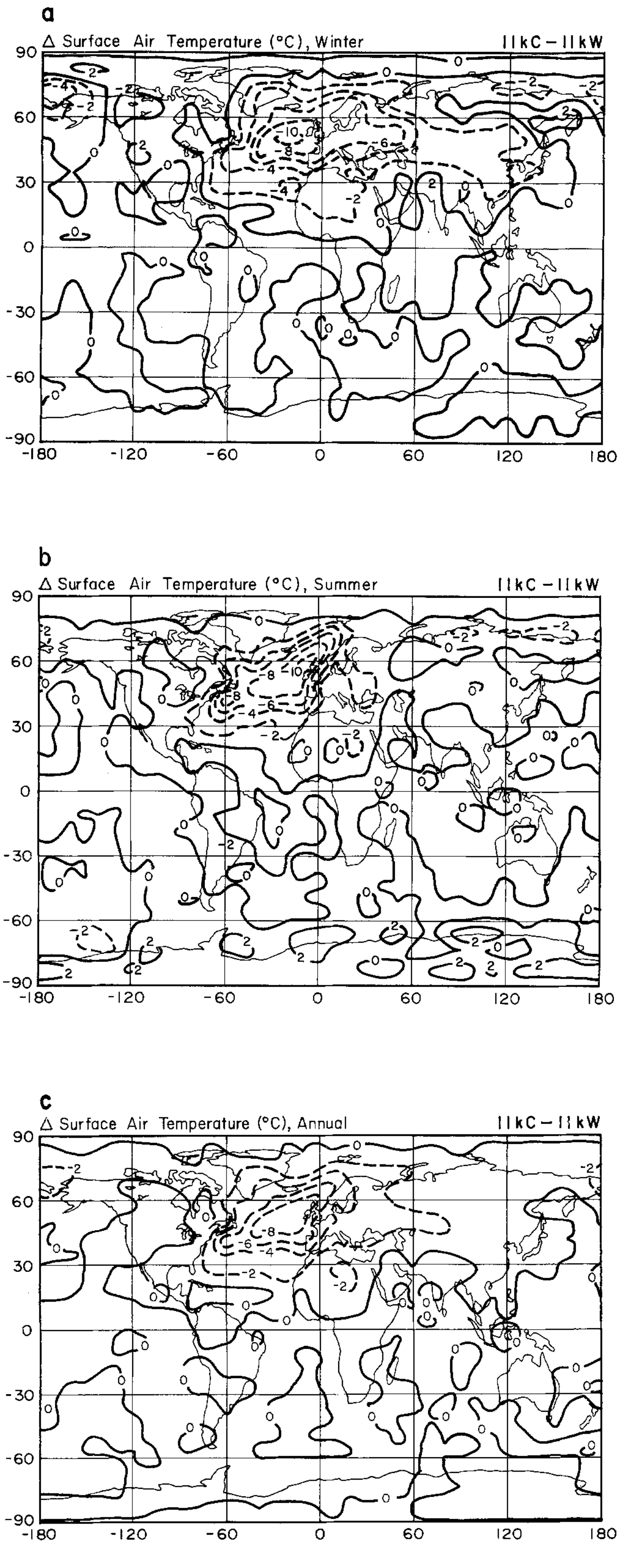


Fig. 12.a Surface air temperature change during winter, 11 kC–11 kW. **b** Surface air temperature change during summer, 11 kC–11 kW. **c** Surface air temperature change, annual average, 11 kC–11 kW

Table 7. Differences in annual average quantities 11 kC–11 kW

	Middle US	North Atlantic	Middle Europe	Siberian plateau	North Pacific
T_s (°C)	−0.1	−5.9	−4.1	−0.7	0.0
P (mm d^{-1})	0.3	−0.7	−0.5	−0.1	0.0
P–E (mm d^{-1})	0.2	0.6	0.0	−0.1	0.0
C (%)	3.0	5.0	−3.0	1.0	−1.0
Net rad (W m^{-2})	−1.0	−2.0	1.0	0.0	1.0
Net heat (W m^{-2})	1.0	35.0	2.0	0.0	−2.0

greater than the precipitation decrease in the region of the Gulf Stream and North Atlantic Drift; again, if runoff changes were small (they are in the model), this would imply a freshening of the North Atlantic.

The results for specific regions are given in Table 7 and are very similar to those for 0 kC–0 kW (Table 6). The annual average temperature change is confined to the North Atlantic and the west through central European region, with less of an effect noted for the Siberian plateau than occurred in the present climate perturbation experiment. These regions also experience precipitation decreases, but as noted, over the North Atlantic evaporation decreases more. The net heat flux into the surface waters increases strongly due to the decreased evaporative heat loss, so again the ocean would have warmed if ocean transports were unchanged. This is shown more clearly in Fig. 15; the change in net heat at the surface is due to the colder ocean temperatures. Note that a deficit in heat input exists to the south and east. For the colder ocean temperatures to have been maintained would have required altered ocean heat transports producing energy divergences of the magnitude shown along the present position of the Gulf Stream and North Atlantic drift. The region to the south and east would presumably have been the recipient of the energy convergence. The shift in Gulf Stream position required is similar to that suggested for 18 k by CLIMAP data, which is to be expected since the 18-k temperatures were used in the Younger Dryas simulation. Thus, a change in ocean transports would seemingly have been necessary to maintain the cold ocean temperatures unless some other cold water source (e.g., continued ice melt) or energy reduction mechanism existed.

It is interesting to note that the presumed Gulf Stream change would resemble the resultant jet stream change (Fig. 13c). As the jet stream location strongly influences the movement of storms and strong surface winds, a shift in the jet stream would conceivably influence the surface wind stress pattern responsible for surface ocean currents. An investigation of the consistency of wind and ocean current changes would require a coupled atmosphere-ocean model, and the magnitudes involved depend on

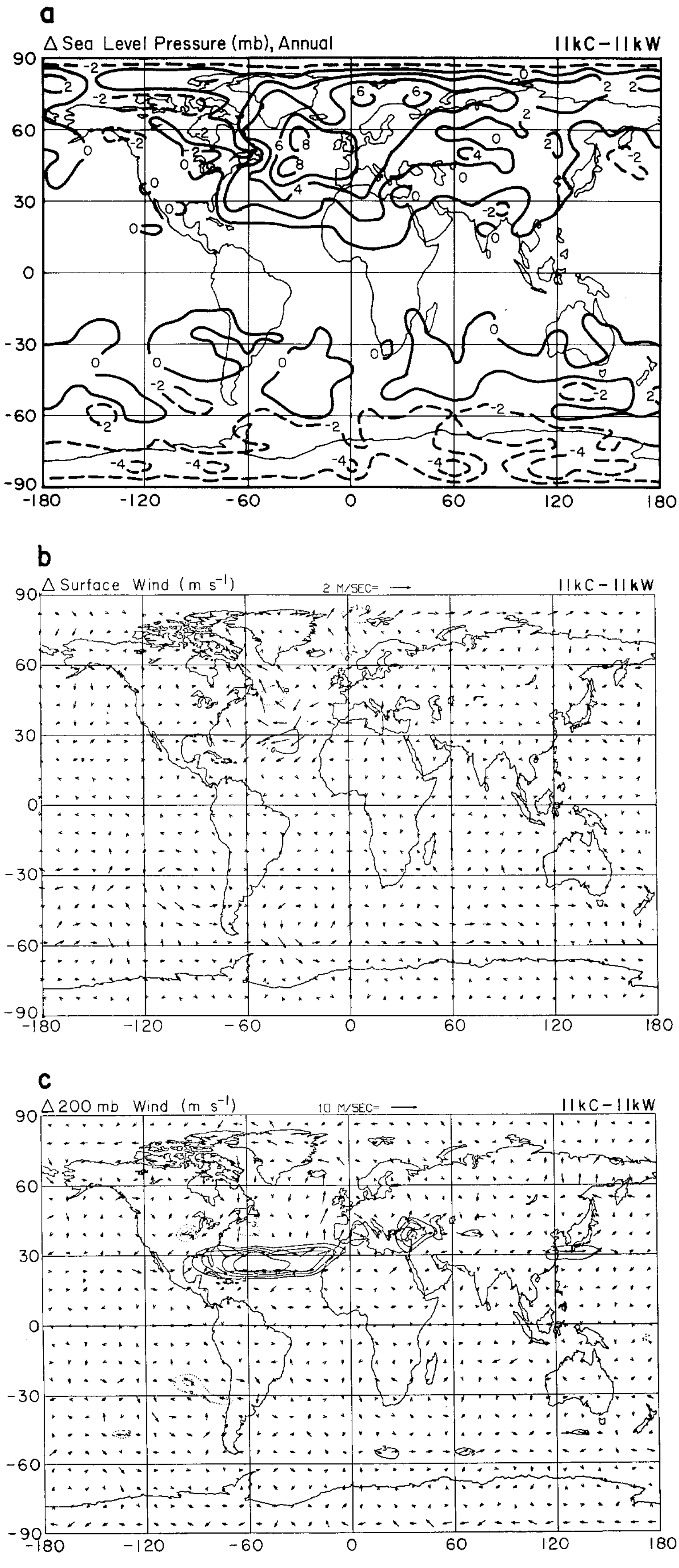
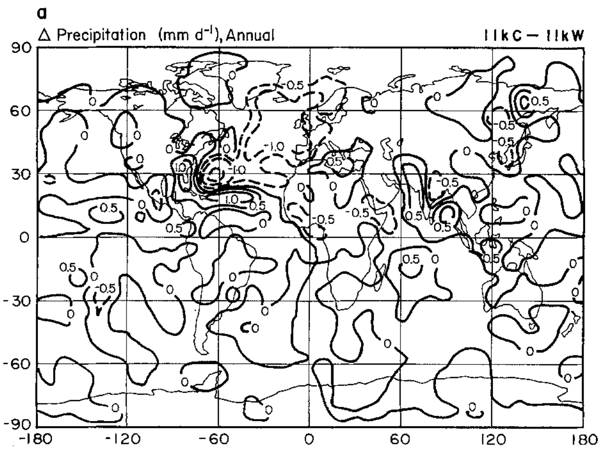


Fig. 13.a Sea level pressure change, annual average, 11 kC–11 kW. **b** Surface wind change, annual average, 11 kC–11 kW. **c** 200-mb wind change, annual average, 11 kC–11 kW



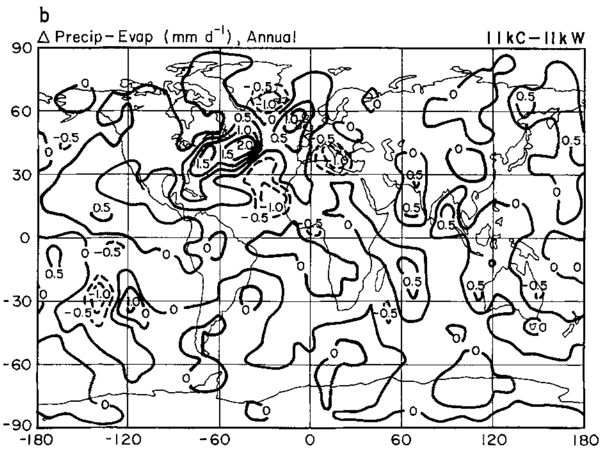


Fig. 14.a Precipitation change, annual average, 11 kC–11 kW. **b** Change in precipitation minus evaporation, annual average, 11 kC–11 kW

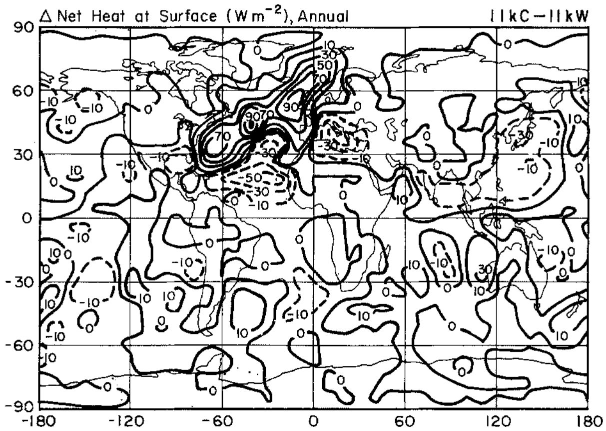


Fig. 15. Net heat at the surface, annual average, 11 kC–11 kW

precisely how cold the Younger Dryas ocean temperatures really were. If the atmospheric dynamic forcing is not consistent, the atmosphere would be continually attempting to drive the Gulf Stream back to its more northward position, implying the Younger Dryas cooling could be maintained only with the aid of additional cooling mechanisms. The results shown in Fig. 15 also indicate that the climate change induced by the cold North Atlantic waters alone would have produced a slight cooling over portions of the North Pacific if those temperatures had been allowed to vary, but the effect would have been small unless ocean dynamical changes or other feedbacks occurred.

The similarity of the results of 0 kC–0 kW and 11 kC–11 kW indicates that the effect of the colder North Atlantic sea surface temperatures is not greatly influenced by boundary conditions such as land ice and orbital parameters. The similarity also increases the confidence in the significance of the results of each experiment.

11 kW–0 kW

Before exploring ways in which the Younger Dryas climate differed from the current climate we need to understand the influence of the change in orbital parameters and land ice. In effect, we need to discern the differences between the Alleröd and the present climate. If one assumes that the sea surface temperatures for the Alleröd were similar to those today (Fig. 5), then the change in the two climates can be estimated by looking at the differences in the two control simulations. The Alleröd had significantly greater land ice and precession was approximately 180° out of phase with current conditions, as the sun was closest to the Earth in Northern Hemisphere summer. Thus, at 11 k Northern Hemisphere mid latitude land surfaces received about 7% more insolation in summer and 7% less in winter compared to today. What effects would these combined changes have on climate?

The surface air temperature change between the two control runs is shown in Fig. 16 (a–c) for winter, summer, and the annual average. In winter the presence of the ice sheets and the decreased midlatitude insolation combine to produce large areas of much colder temperatures. In summer two effects are seen; the increased summer insolation provides for a warmer climate over Asia and North America, in close agreement with the results of Kutzbach and Guetter (1984) for a 9-k climate simulation. However, the large warming indicated for eastern Europe during 11-k summer in this model is completely missing in their reconstruction. This difference is not the result of the slightly different time periods – the insolation change between 9 k and 11 k is small. Instead, it seems to be related to the existence of the ice sheets, which were not present at 9 k.

In estimating the climate sensitivity implied by the full ice age climate, Hansen et al. (1984) ran a series of experi-

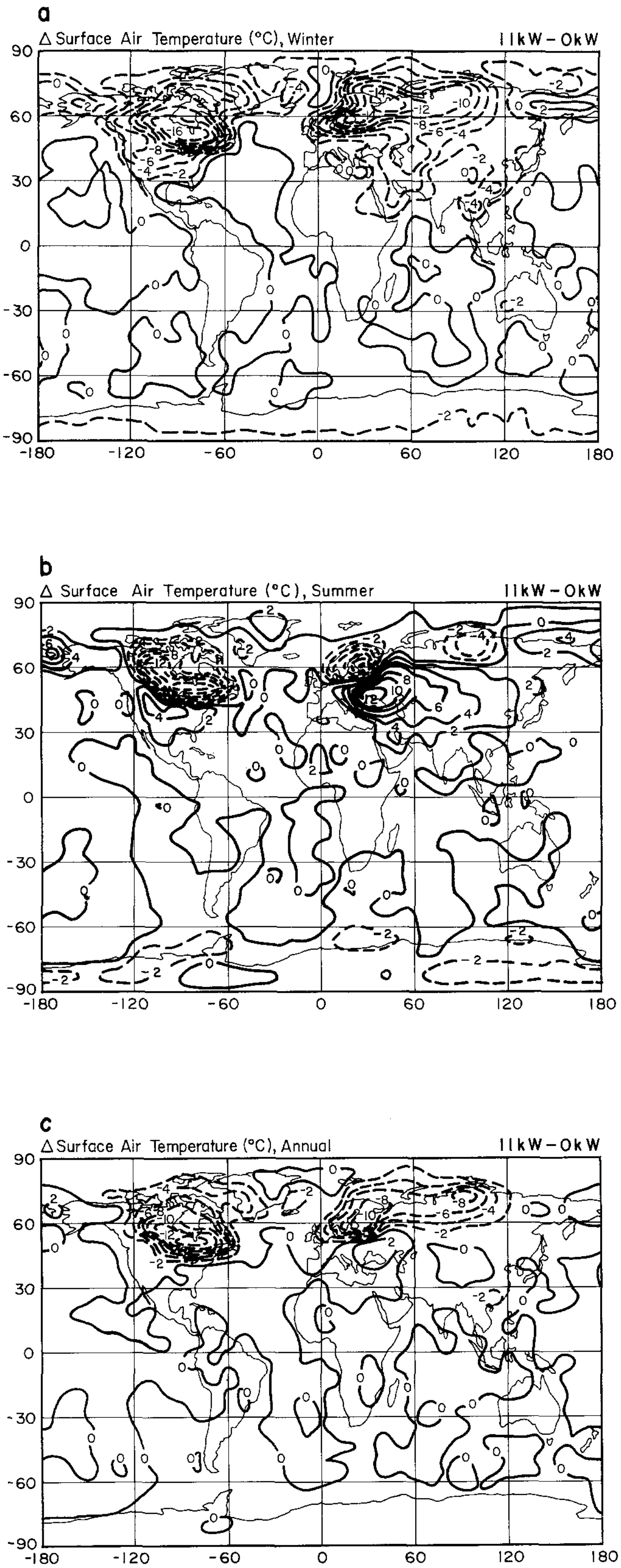


Fig. 16.a Surface air temperature change during winter, 11 kW–0 kW. **b** Surface air temperature change during summer, 11 kW–0 kW. **c** Surface air temperature change, annual average, 11 kW–0 kW

ments to isolate the effects of the ice sheets. In one experiment, the ice sheets were reduced in elevation to 10 m. Compared to the climate simulation using the CLIMAP (1981) ice sheet topography, the results show the elevated ice sheets produced a warming of some 10°C in eastern Europe during summer. Apparently, subsidence of air from over the higher ice sheets is producing warming during this season in regions to the south and east. The effect is absent in winter, or at least masked by the overall cooling induced by the ice sheets. The similarity of these results to the temperature changes in this area which occur in the 11-k experiments implies that it is the ice sheet effect which is responsible for a good part of the summer warming from 11 k to 0 k over Europe. It is thus necessary to provide an accurate specification of the ice sheet topography during the late-glacial in order to the reproduce the proper magnitude of both cooling above the ice sheet and warming near its perimeter. Note that both the orbital variations and the ice sheet effects are likely minimized somewhat due to the specification of the sea surface temperatures at the same values for both runs; the minimizing effect is most noticeable near the coasts and in the Southern Hemisphere, where little temperature change is noted.

The presence of the Laurentide ice sheet provides for increased precipitation to its south and east (Fig. 17a). The temperature contrast in the vicinity of the ice sheet edge increases storm formation there (as indicated by the change toward increased cyclonic or counterclockwise rotation of the surface wind in Fig. 17b); it also increases the jet stream which helps direct storms tracks along that path (Fig. 17c). The cyclonic curvature of the jet-stream-level winds over eastern Canada is indicative of an upper level trough which has formed above the Laurentide ice sheet due to its cold temperatures. This is consistent with increased storm formation and precipitation over eastern North America, and decreased precipitation to the west (Fig. 17a), in accordance with the suggestion of Manley (1971) in explaining the northwestward swing of the margin of the icesheet through Minnesota and beyond. In addition, the strong jet stream zonal flow south of the ice sheet helps establish the predominance of Pacific maritime air over the eastern United States, limiting the winter

Table 8. Differences in annual average quantities 11 kW–0 kW

	Middle US	North Atlantic	Middle Europe	Siberian plateau	North Pacific
T_s (°C)	−0.7	0.0	−0.3	−2.6	0.0
P (mm d^{-1})	−0.2	−0.1	−0.6	0.2	0.0
P–E (mm d^{-1})	0.0	−0.3	−0.2	0.1	0.2
C (%)	−4.0	2.0	−6.0	0.0	1.0
Net rad (W m^{-2})	3.0	1.0	9.0	4.0	2.0
Net heat (W m^{-2})	0.0	−3.0	2.0	3.0	10.0

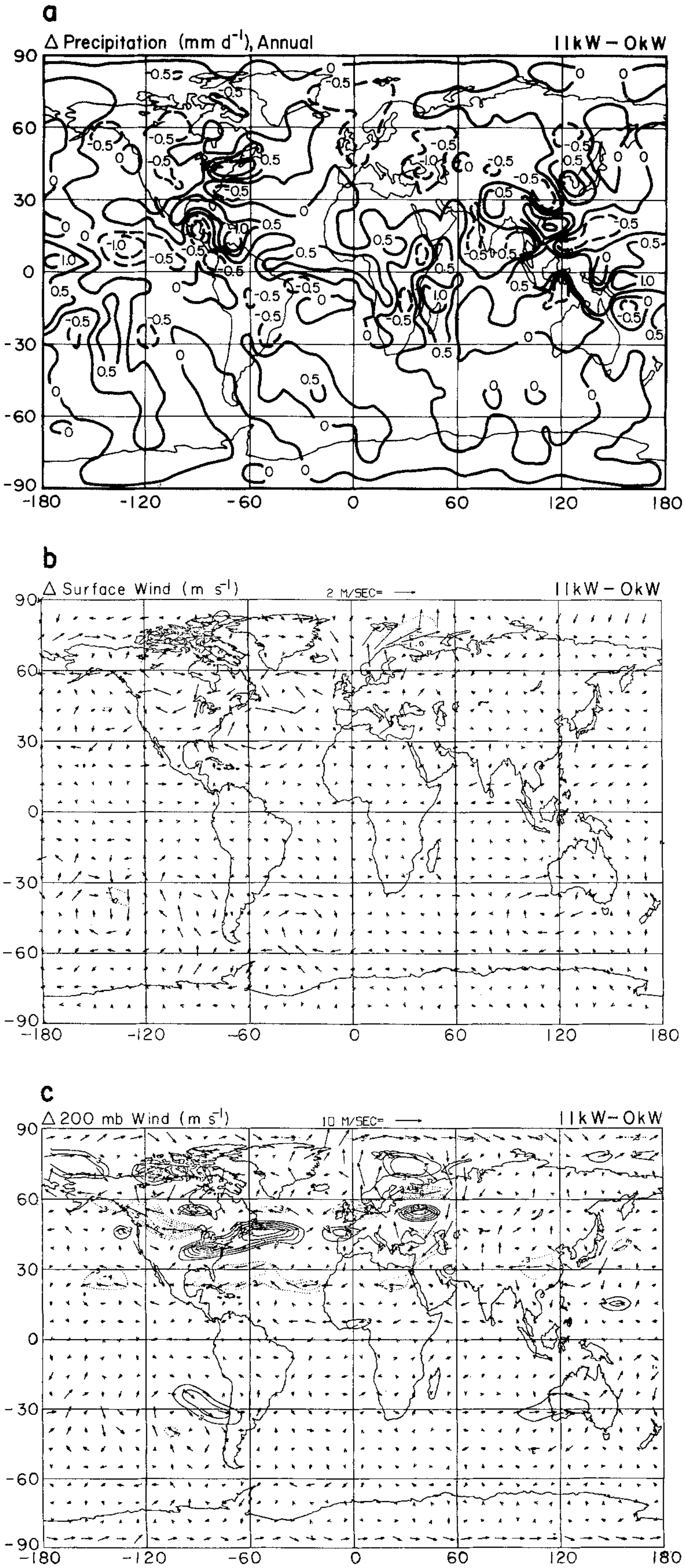


Fig. 17.a Precipitation change, annual average, 11 kW–0 kW. **b** Surface wind change, annual average, 11 kW–0 kW. **c** 200-mb wind change, annual average, 11 kW–0 kW

cooling in this region (Fig. 16a), in agreement with the analysis of Delcourt and Delcourt (1984).

Results for specific regions are given in Table 8; the annually averaged temperature depression in the continental interiors represents the difference between summer warming and winter cooling due to the changes in precession, as well as the influence of ice sheets. In addition, the summer warming over Europe is associated with reduced cloud cover and thus more radiation is absorbed at the surface.

11 kC–0 kW

How different was the Younger Dryas from our climate today? Differences result from increased land ice, changed orbital parameters, and altered North Atlantic sea surface temperatures. The changes in surface air temperature for winter, summer, and the annual average are shown in Figs. 18 a–c. In winter all three phenomena are acting in concert (Figs. 6a or 12a and 16a) to produce an extensive area of cooling throughout mid and upper latitudes in the Northern Hemisphere. In summer, the orbital change, ice sheet influence, and the reduced downstream advection characteristic for that season result in warming of the midlatitudes of Europe, Asia, and North America, as well as in Alaska, despite the colder North Atlantic sea surface temperatures. Precipitation was reduced from current values over the colder Atlantic ocean waters, but greater in the region just to the south (Fig. 19a). This difference as well as the surface wind and jet stream changes (Fig. 19b, c) are effectively composites of the changes due to the ice sheets (Fig. 17) and the cold North Atlantic temperatures (Fig. 10). For example, the colder Atlantic shifts the upper air trough eastward from its position in eastern Canada during the Alleröd simulation (11 kW), extending the strong west to east flow across the ocean (cf. Figs. 19c, 17c).

The values for the different regions are shown in Table 9. The results are the sum of Tables 6 and 7, and are thus consistent with the explanations given previously. Note that for 11 kC–0 kW more regions experience significant cooling than in the other comparisons and that

Table 9. Differences in annual average quantities 11 kC–0 kW

	Middle US	North Atlantic	Middle Europe	Siberian plateau	North Pacific
T_s (°C)	−0.8	−5.8	−4.4	−3.3	0.0
P (mm d^{-1})	0.1	−0.8	−1.1	0.1	0.0
P–E (mm d^{-1})	0.2	0.3	−0.2	0.0	0.2
C (%)	−1.0	7.0	−9.0	1.0	0.0
Net rad (W m^{-2})	2.0	−1.0	10.0	4.0	3.0
Net heat (W m^{-2})	1.0	30.0	4.0	3.0	8.0

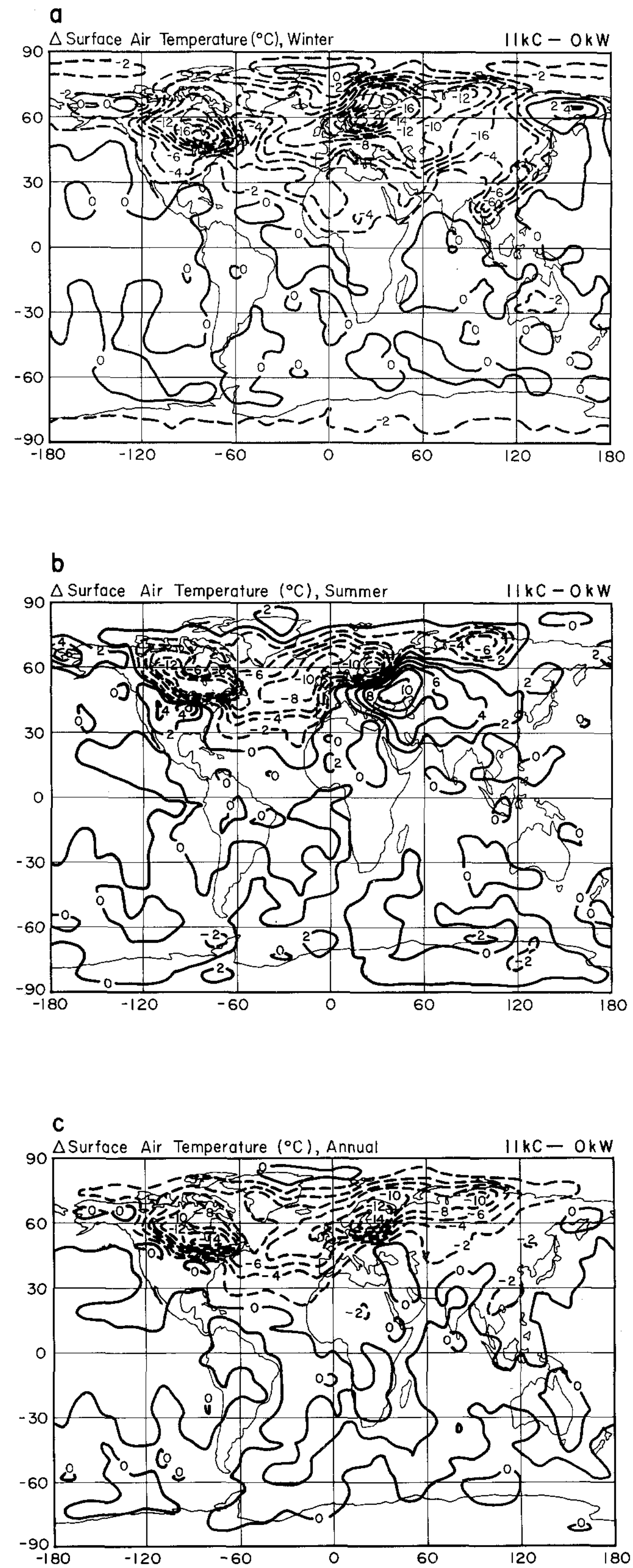


Fig. 18.a Surface air temperature change during winter, 11 kC–0 kW. **b** Surface air temperature change during summer, 11 kC–0 kW. **c** Surface air temperature change, annual average, 11 kC–0 kW

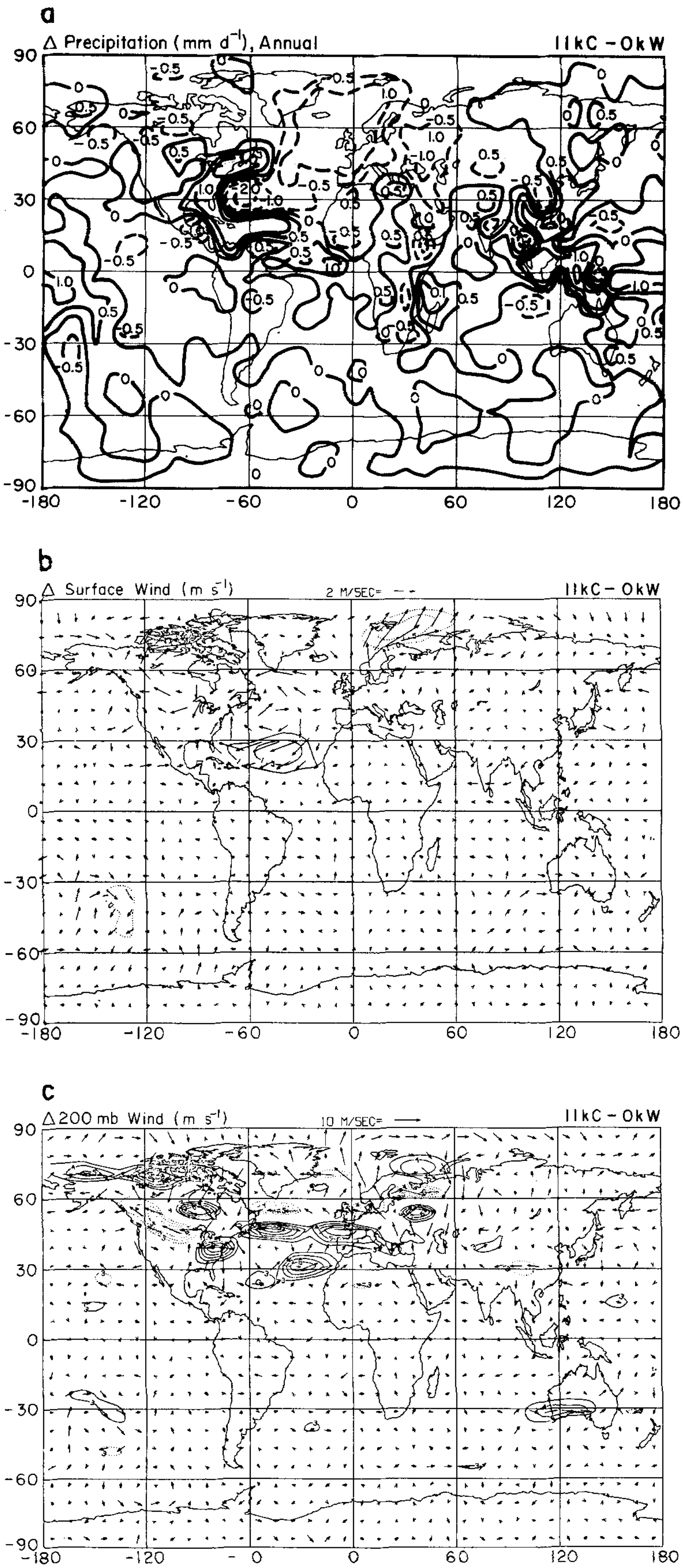


Fig. 19.a Precipitation change, annual average, 11 kC–0 kW. **b** Surface wind change, annual average, 11 kC–0 kW. **c** 200 mb wind change, annual average, 11 kC–0 kW

both the North Atlantic and North Pacific regions would warm if the sea surface temperatures were allowed to adjust.

We now summarize the results by addressing the questions raised in the Introduction. First, what is the effect in the model of the colder ocean temperatures in the North Atlantic, and how do these results compare with paleoclimatic evidence? The influence of the colder North Atlantic sea surface temperatures on the surface air temperatures in the two different experiments (0 kC–0 kW, Figs. 6-8; 11 kC–11 kW, Fig. 12) is nearly identical. This indicates that in the model the effect of colder North Atlantic waters on climate is essentially independent of the insolation and land ice distributions. Changes in additional fields (sea level pressure, precipitation, wind; Figs. 9–11, 13, 14) are also very similar. Thus, for comparison with paleoclimatic evidence, we could use either experiment, although the 11-k orbital parameters and land ice provide more appropriate boundary conditions. We reiterate that the 11-k experiments represent a crude simulation of the Alleröd/Younger Dryas, for which a better specification of sea surface temperatures, sea ice, land ice, etc., is needed.

The change in surface air temperature produced in the model in western Europe generally agrees with the pattern and magnitude (i.e., ca. 7°C) of temperature change deduced from paleoclimatic investigations (Figs. 3, 4; Tables 1–4). The model-simulated cooling is most intense in western Europe and decreases toward Asia. The eastward extent is greater at higher latitudes, but ambiguity exists as to the eastward extent of a Younger Dryas vegetational response. (Figs. 3, 5). Note that the annual average cooling over the Siberian plateau in 0 kC–0 kW (Table 6) is greater than that in 11 kC–11 kW (Table 7). Furthermore, it is important to understand exactly which season influences vegetation most; as shown above the effect of the North Atlantic sea surface temperature anomaly is felt further eastward (downstream) in the winter season when the zonal wind field is stronger (Fig. 12a, b). Pollen results from the USSR are important in this regard, and a more comprehensive spatial analysis of the Younger Dryas expression throughout this vast region would be useful.

The southern and western limits of the Younger Dryas cooling depend upon the magnitude and pattern of cooling of the sea surface temperatures. The model indicates cooling would have been felt near the Atlantic as far south as Portugal, and maybe even in extreme northwestern Africa. Terrestrial paleoclimatic evidence for these areas is scarce. To the west, the results show that cooling occurs along the east coast of North America. Palynologic (and lithologic) evidence indicates that some cooling occurred in eastern Canada (Fig. 4). As indicated, the anomalies for the model experiments were determined from the difference between the ice age and current values down to 25°N throughout the North Atlantic, and the relevance of the model-determined cooling results to the west and south depends on how representative these sea surface temperatures are of the actual Younger Dryas values. Further westward over interior North America the model indicates no cooling from the North Atlantic anomaly and the palynological evidence is generally consistent (Table 1).

At all other locations – over Japan, throughout the Pacific, along the west coast of North America, and in the Southern Hemisphere – the model indicates that no statistically significant cooling should have occurred. There is also no indication of significant changes in the hydrological cycle. Where paleoclimate evidence indicates an Alleröd to Younger Dryas change at other locations, the model results suggest that it could not have been caused by changes in North Atlantic sea surface temperatures alone. For example, the description in Table 3 and Fig. 5 of possible Younger Dryas effects in Africa, western North America and the Southern Hemisphere are completely unexplained by the model results. As indicated by the jet stream change (Figs. 10b, 13c) there is an indication that the colder North Atlantic would have influenced the long wave pattern in the Northern Hemisphere which might have initiated changes in weather and temperature patterns at other locations. In fact, there is some verification of the suggestion of Manley (1971) for increased westerly flow (at least at upper levels) over Japan (Fig. 13c). The model results indicate that such effects would have minimal impact on the surface air temperature unless the sea surface temperatures in these other regions were also allowed to change, due, for example, to ocean circulation and/or atmospheric CO_2 concentration changes.

Broecker et al. (1985) suggested that the Pacific Ocean might have become saltier. Currently, there is moisture divergence from the atmosphere over the North Atlantic and moisture convergence over the North Pacific (Peixoto and Oort 1983). If the colder North Atlantic reduced the atmospheric moisture divergence, the convergence over the North Pacific might also decrease, leading to an increase in North Pacific salinity. The model results, however, show no decrease in rainfall over the North Pacific due to this effect, and no decrease in P-E (Tables 6, 7). However, if reduction of North Atlantic Deep Water production led to warmer sea surface temperatures in the North Pacific, evaporation would likely have increased in that region, and altered the moisture balance into the ocean. A sea surface temperature reconstruction for the North Pacific from radiolarians (Heusser and Morley 1985) shows higher summer temperatures in the lateglacial compared with those during the Holocene. This further emphasizes the need to include appropriate global sea surface temperature distributions in a full simulation of the Younger Dryas.

What was the magnitude of the temperature change, both annually and seasonally between the Younger Dryas and today? Temperature changes indicated by paleoclimatic observations and model results are shown in Table 4. As discussed above, the model values are in good agreement with the empirical evidence for the regions surrounding the North Atlantic. The Younger Dryas simulation also emphasizes the importance of the increased seasonality of insolation which characterized the Northern Hemisphere during the late-glacial – Holocene interval. In addition, the model produces amplified cooling/warming in the vicinity of the ice sheets. Denton and Hughes (1983) proposed that summer insolation is the primary control on ice sheet dynamics through its effects on marginal wastage. Yet, despite the substantial summer insolation at mid and high latitudes, 11–10 k paleoclimatic evidence indicates a sharp reversal in the late-glacial warming trend, including glacial growth in the Scottish Highlands, Scandinavia, and the USSR. The model produced increases in winter snow depth of up to 50% in northwestern Europe despite reduced winter precipitation. Thus, the model results are not inconsistent with the paleoenvironmental data which indicates glacial growth during this interval, and they emphasize the importance of the cooling that must have occurred in these regions.

In response to the greater extremes in insolation and the ice sheet presence, the model produces warmer summers and colder winters at Northern Hemisphere midlatitudes (changes of 5°–10°C in each season). This increased seasonality has implications for interpretation of the late-glacial record in the midcontinental United States and eastern Europe, where no-analog assemblages of boreal and temperate species characterize the pollen record. For example, the absence of trees has been used to infer low summer temperatures. However, trees may be limited by other features of a more continental climate which result from increased seasonality, such as very cold winters, severe wind exposure, severe spring thaws and floods, and highly unstable soils (Godwin 1975). Under these conditions, the landscape may have been devoid of trees even with warmer summers than those prevalent today. In addition, increased seasonality would have affected plant reproduction, competition, and migration. Several authors (Starkel 1977, Sissons 1979; Van Geel and Kolstrup, 1978; and Van Zeist and Bottema 1982), among others, conclude that a greater annual temperature range did exist in Europe and the USSR (Khotinskiy 1983) in the late-glacial period. Records of vegetational change, beetle assemblage change, permafrost distribution, and glacial evidence all indicate this enhanced seasonality.

Wright (1985) raises the possibility of reduced seasonality to account for some of the no-analog, late-glacial assemblages of temperate and boreal species, suggesting that if summers were cooler and winters milder, temperate species could have occupied regions farther north. Such assemblages have been observed in the midlatitudes (e.g., Minnesota, Amundson and Wright 1979). Wright suggests that the influence of the North American ice sheet in blocking cold Arctic air outbursts may have led to milder winters, and the proximity of ice to cooler summers. The model indicates that in regions close to the ice warmer summer temperatures were possible (Fig. 18b), and winters were more severe. Reexamination of late-glacial sites with a better understanding of species tolerances to temperature ranges is needed to decipher the palynological records of climate change that characterized the Alleröd-Younger Dryas fluctuation.

Mammals would also be subject to the effects of increased seasonality. The fossil records reveal numerous extinctions of large animals during the late-glacial, especially during the 11–10 k interval (Lundelius et al. 1983; Sempken 1983). The cause(s) of these extinctions has long been debated (human overkill versus climatic change hypotheses). The effects of increased seasonality, enhanced by the ice nearby, would likely have had marked influence on animal distributions. Individual species tolerance for winter and summer temperatures would affect reproduction and migration, as well.

What are the implications for the future, as climate warms due to increased CO_2 and other trace gases? This obviously depends on the cause(s) of the Younger Dryas event: without knowing the reason(s) for the sudden southward shift of the oceanic polar front at 11 k it is difficult to assess its relevance for understanding future climate change. If it required large amounts of ice at high latitudes to provide for ice melt, as existed during the Alleröd and the other periods of possible rapid oscillations (Broecker et al. 1985), then such an effect would be unlikely under current conditions. Furthermore, current summer insolation at high latitudes is low in comparison with that at 11 k. No full event of this type has occurred during the past 10,000 years with the present Greenland ice volume; however, there has been no warming to match that projected for the next century.

If, for example, a reduction in North Atlantic deep water production were to occur, and if it led to substantially colder North Atlantic sea surface temperatures, then future warming over Europe would be retarded or eliminated. The magnitude of the temperature changes which result from the colder North Atlantic sea surface temperature influence on the present climate (0 kC–0 kW) are similar to those estimated from doubled CO_2-induced warming (e.g., Hansen et al. 1984).The apparent variability of the North Atlantic has the potential to impact the coming climatic warming, but we are not yet in a position to estimate its likelihood.

What is needed to enhance our understanding of the Younger Dryas sudden cooling? Further research is necessary to provide answers to the following questions:

What was the geographical distribution of the event – was the Younger Dryas a worldwide phenomenon or was it confined to the vicinity of the North Atlantic (Fig. 5)? What were the magnitudes of temperature changes over the continents? What were ocean temperatures like in other parts of the world? What magnitude of ice melt and fresh, cold water intrusion into the North Atlantic is needed to reduce North Atlantic deep water production or cool North Atlantic sea surface temperatures substantially? How relevant are the other possible periods of rapid oscillations as indicative of the tendency of the system to exhibit such transitions; do they indicate the boundary conditions (e.g., land ice volume and summer insolation) that are necessary for rapid changes to occur? We need to investigate whether events such as the Younger Dryas are characteristic features of our climate system.

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